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UNION GÉODÉSIQUE ET GÉOPHYSIQUE INTERNATIONALE

INTERNATIONAL UNION OF GEODESY AND GEOPHYSICS

ASSOCIATION INTERNATIONALE
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INTERNATIONAL ASSOCIATION
OF SCIENTIFIC HYDROLOGY

ASSEMBLÉE GÉNÉRALE DE HELSINKI

25-7 — 6-8 1960

GENERAL ASSEMBLY OF HELSINKI

COLLOQUE SUR LA GLACIOLOGIE ANTARCTIQUE

SYMPOSIUM ON ANTARCTIC GLACIOLOGY

PUBLIÉ AVEC L'AIDE FINANCIÈRE DE L'UNESCO

PRIX : 150 Frs belges

PUBLICATION N° 55

DE L'ASSOCIATION INTERNATIONALE D'HYDROLOGIE SCIENTIFIQUE

SECRÉTAIRE : L.J. TISON
BRAAMSTRAAT 61, (RUE DES RONCES)
GENTBRUGGE (BELGIQUE)
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ETUDE DE L'ACCUMULATION DE TERRE ADELIE

C. LORIUS

RÉSUMÉ

Des mesures directes (balises d'accumulation) complétées par des études stratigraphiques dans les couches superficielles de névé ont été réalisées au cours de l'AGI dans le secteur de Terre Adélie. Les résultats indiquent une nette décroissance de la quantité de neige accumulée lorsque l'on s'éloigne de la côte vers l'intérieur du continent antarctique; on exprime une relation théorique entre l'accumulation et différents facteurs naturels (distance à la côte, pente) utilisant les données des raids récents.

Le problème du datage par les méthodes classiques de la stratigraphie est discuté.

TENEUR EN DEUTERIUM DE PRECIPITATION DANS L'ANTARCTIQUE APPLICATION AU PROBLEME DU DATAGE DES COUCHES DE NEVE

C. LORIUS

RÉSUMÉ

Les analyses effectuées par spectrométrie de masse sur 20 échantillons de précipitations prélevés au cours de l'été antarctique 1959-1960 en Terre de Victoria montrent une relation pratiquement linéaire entre leur teneur en Deuterium et leur température de formation estimée à partir de radiosondages dans l'atmosphère. Cette relation est en bon accord avec le calcul théorique et les mesures effectuées en laboratoire.

Par suite des différences de température de nuages donnant les précipitations entre les périodes d'été et d'hiver, la teneur en Deuterium des différentes couches saisonnières du névé devrait être différente; on en trouve à B 61 (71°08′S, 139°11′E) un bon accord entre le datage obtenu par la méthode classique d'étude (stratigraphie, résistance à la pénétration, densité) et le profil obtenu à partir des concentrations en Deuterium des différents niveaux. L'utilisation de cette méthode à l'important problème de glaciologie des accumulations annuelles semble donc possible.

THE US-IGY CONTRIBUTION TO ANTARCTIC GLACIOLOGY

R.L. CAMERON and R.P. GOLDTHWAIT

SUMMARY

Snow accumulation studies made at the base stations give the following water equivalent values for average annual accumulation during more than a decade: Little America, 21 cms; Marie Byrd, 18 cms; South Pole, 7 cms; an Wilkes (S-2) 13.3 cms. These values, with the traverse determinations, represent the accumulation over a third of Antarctica. An accumulation map of Antarctica is presented, using these data and the data collected by other nations. The largest accumulation is on the east side of the Weddell Sea, and the smallest accumulation is in the area of the Pole of Inaccessibility.

A map of the mean annual air temperature for Antarctica shows that the «cold pole» (–57° c to –59° C?) is located in the vicinity of Vostok II. The mean annual air temperature for both Little America and Ellsworth is –24 °C, and the mean annual air temperature for the coast of east Antarctica ranges between –9 °C and –11 °C.

Relative motion networks established at inland stations in 1958 and resurveyed in 1959 have not shown appreciable expansion. Measurement of absolute movement (2.1 m/day) of the Vanderford Glacier, near the Windmill Islands, indicates that glaciers along the coast of east Antarctica may be discharging ice at a greater rate than previously postulated. Skelton Glacier is moving into Ross Ice shelf at 0.28 m/day.

Observations in the Windmill Islands, Cape Hallet area, Sentinel Mts., and the Dufek Massif further confirm the greater thickness of the Antarctic ice sheet at an earlier stage. The Dufek Massif, 200 miles from the coast, has a large ice free valley containing several recessional moraines. Not only has the ice thinned in this area, but the present relationship between low precipitation and relatively high summer temperatures maintains this ice- and snow-free valley.

1. INTRODUCTION

The total contribution made by the United States to the glaciology of Antarctica during the IGY is far too big to summarize in one article. Actually the contribution is expressed by several other papers given at the 1960 I.U.G.G. Congress in Helsinki and will be presented in many other analytical papers. The first effort at The Ohio State University was in the reduction of routine data from all U.S. stations and traverses in Antarctica. Inevitably, as this was done, certain general patterns and facts emerged. These preliminary general observations are presented here.

Some detailed studies, made elsewhere but financed at The Ohio State University as part of the total effort, have already reached final analysis (Péwé: glacial geology of Taylor Dry Valley, Zumberge: deformation of Ross Ice Shelf, and Hoinkes and Dalrymple: micrometeorology of some antarctic stations). These experts are developing special interests and objectives. However, there are many more subjects for analysis to be found in the hundreds of pages of data. Study has already begun on annual snow layers and on problems of glacial geology, but there remains untouched a wealth of study concerning compaction laws and processes, crystal area versus depth, heat flow in the firn and ice, and related problems.

2. NET ACCUMULATION

The main objective of the glaciological program in Antarctica during the IGY was the study of the regime of its vast ice sheet. The regime, negative, positive, or in equilibrium, is a function of net accumulation and net loss. The most readily deter-

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mined factor is the net accumulation, whereas the net loss is more difficult to assess. Snow accumulation at six U.S. bases was measured with a series of stake nets. The average annual accumulation in centimeters of water for these bases is as follows: Little America, 21 cm; McMurdo Sound, 18 cm; Marie Byrd, 18 cm; Ellsworth, 20-25 cm; Wilkes inland station (80 km from coast), 13 cm; and the South Pole, 7 cm. The various oversnow and airborne traverses made a total of 290 snow pit studies in which the annual accumulation was determined and these studies reveal the distribution of accumulation over West Antarctica. These values represent the accumulation over one-third of Antarctica and when combined with the values obtained by other nations it is possible to prepare an accumulation map of Antarctica (Fig. 1).

The lines of equal net accumulation are 5, 10, 20, 30, and 40 centimeters of water. It is herewith proposed that such lines of net accumulation be called isohydropleths. The distribution of accumulation over Antarctica, as represented on this map

Fig. 1 — Accumulation map of Antarctica.

by the isohydropleths, is of necessity preliminary as it is based on only 300 known points.

It is interesting to note the probable extent of the area which retains less than 5 cm of water per year. Measurements at Sovetskya show as little as 2.5 to 3.5 cm per year. The area between the Ross and Filchner ice shelves receives 10 to 20 cm of water per year. The greatest accumulation is along the eastern side of the Weddell Sea from Halley Bay, 26° 36′ west longitude, to Norway Station at 2° 52′ west longitude, where the accumulation ranges from 34 cm to 49.5 cm, respectively. Almost the entire coastal area of East Antarctica from the Belgian Base nearly to Cape Adare is within the limits of the 10 and 20 centimeter isohydropleths. However, high accumulation values have been obtained at Mirny 60-70 cm, and at Dryglaski Island 70-80 cms, and low annual *precipitation* is recorded from Davis Base, 6,5 cm. Thus there are localities along the coast of East Antarctica which have anomalous accumulation as a result of topographic irregularity or increased maritime influence.

The total amount of annual net accumulation for the continent is obtained by determining the area bounded by adjacent isohydropleths, multiplying this figure by the average accumulation for that interval, and then adding these increments. The result is a net accumulation of 1.7×10^{18} gm/yr. Mellor (1959, p. 532) in an article on the mass balance of Antarctica, concluded from a study of meridional profiles of accumulation obtained by Shumskiy, Schytt and A.N.A.R.E. that the annual net accumulation for the continent is 1.7×10^{18} gm/yr. It is noteworthy that identical results are obtained from different computational procedures.

3. Ice motion

The most important factor in ascertaining the mass balance of the Antarctic is the amount of ice being discharged by calving along the periphery of the continent. The U.S. scientists have measured the rate of motion of one such ice stream discharging from Antarctica. This glacier, the Vanderford Glacier near Wilkes Station, is moving at the rate of 2.1 m/day. In this same area, ice reaching the coast in a broad front and not channelled is moving at the rate of 15 cm/day.

The rate of motion of the Skelton Glacier moving into the western side of the Ross Ice Shelf, just south of McMurdo Sound, is 0.28 m/day.

Relative movement studies have been made at several of the inland stations, but analysis is not yet complete.

4. Mean annual air temperatures

At all stations and on all traverses, temperature measurements were taken at a depth of 10 m to ascertain the mean annual air temperature. The results of nearly 300 observations are plotted along with the values supplied by other nations (Fig. 2). The mean annual isotherms are −10°, −20°, −30°, −40°, −50° and −55°C. The map is preliminary and will be modified as new data are obtained. The temperature values used in plotting the map were obtained from the records of the U.S. stations and traverses and from a temperature map compiled by the Scott Polar Research Institute. At this time it is not considered appropriate to plot the 5° isotherms except for the −55°.

This diagram shows that the «cold pole» of the Antarctic is where the mean annual air temperature is between −55° and —57°C. Both Russian inland sites, Sovetskaya and the Pole of Relative Inaccessibility are within this area. Zakiyev *in* Bugayev (1960, p. 32) notes that between these two stations an area exists with slightly higher

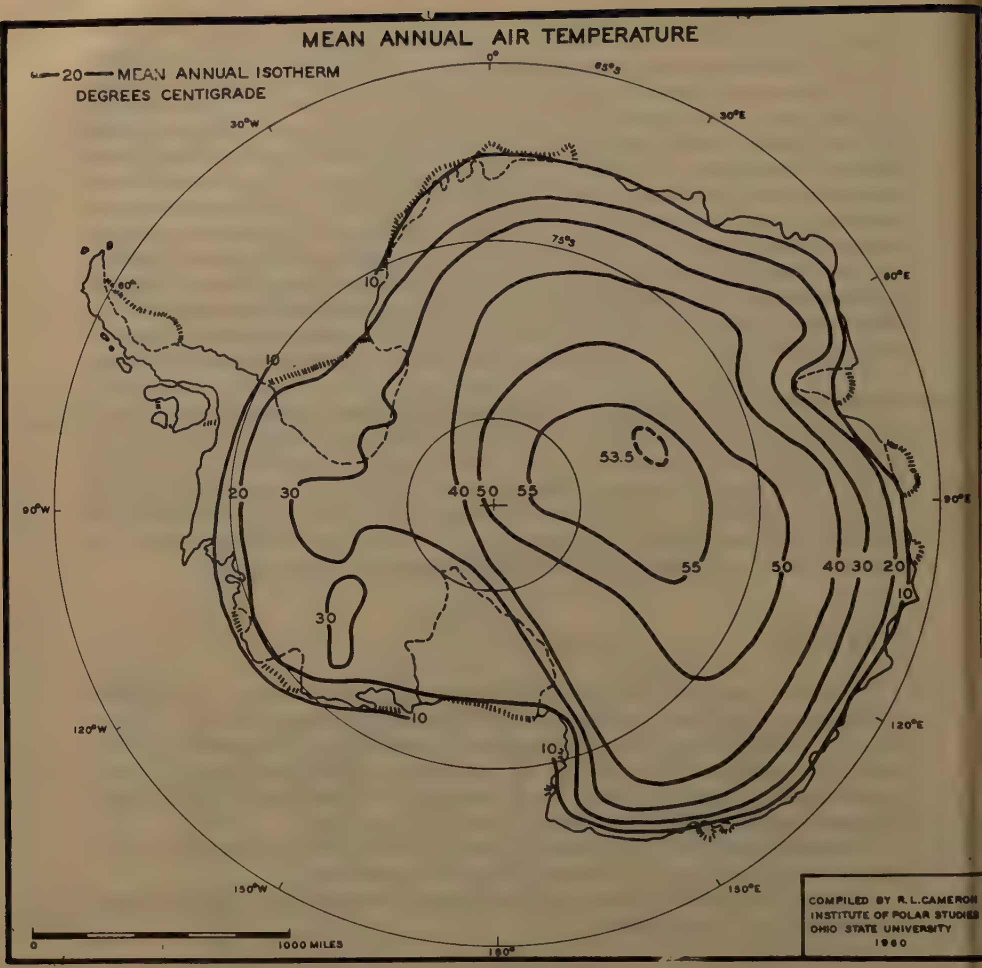

Fig. 2 — Mean annual air temperature map of Antarctica.

temperatures (—53.5°), but gives no explanation for such an increase. This higher temperature zone seems incongruous, but if it actually exists and is not the result of instrumental error, then it would indeed be advisable to study it further for an explanation of this anomaly. It may very well be that the 55° isotherms should lie south of this area.

The Ross and Filchner ice shelves have similar temperature regimes which will facilitate the glaciological comparison of these floating ice bodies. Both shelves lie within the —20° and —30° centigrade annual isotherm. Little America and Ellsworth stations, on the eastern side of their respective shelves, record —24°C as their mean annual air temperature.

5. Greater ice thickness

Considerable evidence has been gathered to corroborate earlier observations of a reduction in the amount of ice coverage in Antarctica. This reduction has not been in coastal areas only, but over the interior of the continent as well. The amount of this reduction varies from one locality to the next, as a result of the relationship of net accumulation and rate of ice flow.

In the Dufek Massif, located 320 km inland of the edge of the Filchner Ice Shelf, Aughenbaugh (1958, p. 183) has recorded a lowering of the general ice surface of 150 to 240 m. A «dry» valley exists in the massif and distinct, well-preserved recessional moraines indicate a present minimum extent of ice and possible present-day retreat of the ice lobes (Figs. 3 and 4).

Fig. 3 — An ice-free valley, above the firn line, in the Dufek Massif. Note the moraines in front of the ice lobes on left. The ice lobes are blue ice and have no firn cover. Note also the polygonal ground. Photo by Paul T. Walker.

Fig. 4 — Aerial view (vertical) of the same ice-free valley shown in Fig. 3. The succession of moraines parallel the ice front. Official U.S. Navy photograph.

Anderson (1958, p. 264) records striations near the Sentinel Mountains on the top of Fisher Nunatak, which rises 183 m above the surrounding ice sheet. Also, in Marie Byrd Land, Doumani (personal communication) has described glacial grooves 450 m up on the flanks of Mt. Sidley (Fig. 5). These grooves are at right angle to the slope.

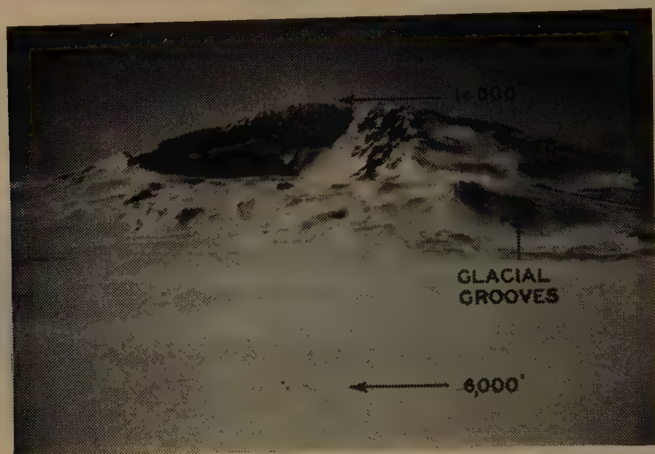

Fig. 5 — Glacial grooves have been described by George Doumani 450 meters up on the flank of Mt. Sidley, Marie Byrd Land. These grooves are at right angles to the slope. Photo by George Doumani.

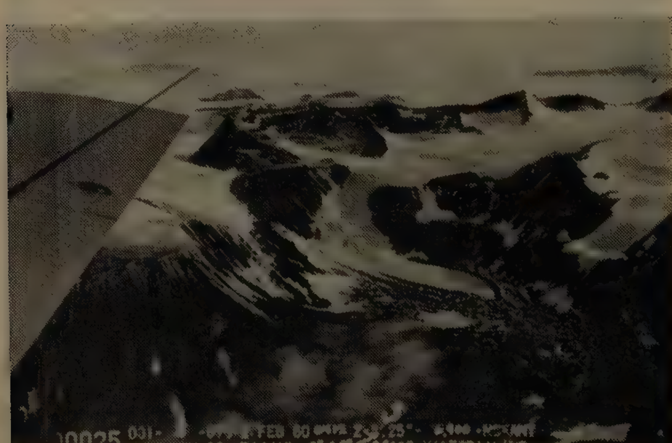

Fig. 6 — Mountains in Victoria Land just south of Rennick Bay showing general ice retreat. Official U.S. Navy photograph.

Mountains in Victoria Land just south of Rennick Bay have empty cirques and shrinking valley glaciers (Fig. 6). Note the recessional moraines. In this area there is a definite negative regime; but whether it can be attributed to decrease of precipitation or to high summer temperatures, or to both of these, is not known.

The Windmill Islands in East Antarctica comprise an area of 75 km², which has been exposed by the retreating ice periphery. Elevated beaches exist as high as 30 m above sea level, and one beach at 23 m contains coralline algae dated with Carbon 14 as 6,040 ± 250 years B.P. (M-1052). This date is a minimum age for the deglaciation of this area.

In the vicinity of McMurdo Sound, Péwé (1959) described several glaciations, and the most recent ice advance, the Koettlitz Glaciation, has a minimum age of 6 000 years.

The few absolute dates available suggest that the present marginal retreat has been maintained for *at least* the last 6,000 years. No absolute dating is presently possible for the inland sites, but the shrinkage there is probably synchronous with that in the marginal areas. The large scale lowering of the surface of the inland ice (about 500 m in some areas) marks a much older event or time interval which is probably a correlative of the retreat of the Ross Ice Shelf from the submarine moraines, and the interval between the McMurdo and Taylor glaciations as described by Péwé in the McMurdo Sound area.

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THE GEOLOGICAL ACTIVITY OF THE ICE COVER IN EASTERN ANTARCTICA

S.A. YEVTEYEV (USSR)

RÉSUMÉ

Les recherches furent effectuées dans la région des travaux de l'expédition soviétique antarctique (78° long. *E* — 110° long. *E*).

La calotte glaciaire antarctique d'une épaisseur pouvant atteindre 3000 m et davantage et possédant une vitesse de progression de près de 1000 m par an dans les secteurs les plus mobiles (écoulements glaciaires), exerce une action intensive sur le lit fondamental. La glace rabote la surface des roches, la burine, entrainant des débris arrachés. La direction du mouvement dans les couches inférieures de la calotte glaciaire varie fortement de place en place d'après le relief du lit. La direction de la hachure glaciaire oscille du NO jusqu'au N et NE.

L'épaisseur de la glace contenant des moraines atteint 100 m dans les écoulements glaciaires. Dans les secteurs moins mobiles de la calotte glaciaire, son épaisseur varie entre 10 et 40 m.

La combinaison des deux mécanismes du mouvement des glaces dans la limite de l'épaisseur contenant les moraines (écoulement visco-plastique et glissement le long des surfaces de rupture interne) concourt à la distribution des débris suivant les axes en long dans le sens de la progression et à donner aux blocs érratiques la forme d'un fer à repasser.

Exerçant son action sur la surface des roches du lit, la calotte glaciaire de l'Antarctide orientale s'efforce d'élargir les vallées, tailler des terrasses sur les versants, augmenter la rugosité du profil en long des vallées.

D'après des calculs approximatifs, une couche de près de 0,05 mm est rabotée annuellement de la surface du lit dans l'Antarctide orientale. Une érosion glaciaire aussi rapide est parfaitement comparable à l'érosion des rivières dans les régions de plaines.

The author limited his investigations to the coast of the continent in the area between 80° east longitude in the west and 110° east longitude in the east. In this area the ice drains to the coast from a vast territory, the highest point of which is in the region of 82° south latitude and 75° east longitude and attains an elevation of more than 4,000 metres above sea level.

The ice is thickest (more than 3,000 metres) 1,000 kilometres from the coast. Closer to the coast it grows thinner at the rate of 1,300 metres per 100 kilometres, then at about 800 metres per 50 kilometres, and is approximately 200 metres thick at the edge of the sheet (near Mirny). The ice sheet thus moves along a very uneven bedding of basic rock. The protuberances and hollows in the rock are reflected in the relief of the surface of the ice sheet not only along the coast, where it is relatively thin, but also in the central regions, where the greatest elevation of the ice sheet coincides with the greatest elevation of the surface of basic rock. The ice breaks away from the continent most actively along the deepest meridionally oriented hollows of the ice sheet bed and here we find the formation of glacial currents which, compared with the adjacent areas of the ice sheet, are deeper and move faster, reaching a speed of up to 1,000 metres a year.

Moving along the badly broken surface of the basic rock, the ice frequently changes its direction. On the surface of the basic rock there may be striping runnig from the north-west to the north-east, and in each concrete case the features of the relief of the rock help to establish why the ice had changed its direction without the need to apply the hypotheses about the changes in the general direction of the spread of the ice in the ice sheet in order to elucidate the presence of striping.

Fragments of rock break away and are imbedded in the body of the ice sheet as

the ice moves along the rock bed. Observations taken where the bottom layers of ice rise to the surface of the ice sheet, a study of the structure of the lower parts of upturned icebergs and seismic data about the thickness of the ice show there is a layer of moraine-containing ice in the base of the ice sheet practically everywhere and, in any case, along the coast. Its thickness varies greatly from place to place. Where the ice sheet is not very broken it is probably not more than 20-30 metres thick. At the base of the ice sheet climbing the Banger hills from the east, the moraine-containing ice has been found to be 40 metres thick. In the glacial currents this thickness increases up to 100 metres.

Moraine-containing ice, which is continuously mixing in the base of the ice sheet, was formed as a result of the exarating activity of the ice itself. The quantity of moraine carried by the ice therefore enables us to judge the intensity of the exarating activity of the ice.

Determination of the percentage of moraine in the ice showed that it distributes unevenly. The greatest quantity of moraine occurs in the low levels of the ice (Table 1).

TABLE 1

Moraine content in the ice crawling up the Banger hills

Level	Depth, metres	Moraine content, %	
		in weight	in volume
1	0-8	0.32	0.11
2	8-20	0.56	0.19
3	20-25	1.21	0.41
4	25-27	3.35	1.19
5	27-30	5.50	2.04
6	30-35	9.60	3.81
7	35-37	23.97	11.84

The edge of the ice sheet crawling over the Banger hills from the east, is a slow-moving sector. Observations in the region of the glacial currents (Table 2) yielded results that were analogous to those given in Table 1.

TABLE 2

Moraine content in the ice of glacial currents

Place of extraction	Situation in cross-section of moraine-containing ice	Moraine content in % of volume
Left edge of Helen glacier	top	0.23
Denman glacier	middle	0.41
Edisto glacier	bottom	10.56

A comparison of the results of determining the percentage of moraine in sections of the weakly differentiated edge of the ice sheet (Table 1) and in the glacial currents (Table 2) gives us grounds for considering that there is an average of about 2 per cent moraine in the moraine-containing ice in any section of the ice sheet.

This activity takes place through the imbedding in the ice of fragments of basic rock broken away from the bed, through the breaking off of the protuberances of the basic bed especially during sudden shifts of the ice along the shear planes, and also through the scratching of the surface of the basic rock by rock fragments in the ice.

Judging by the relief of the ice-free sections of the surface, this exarating activity of the ice leads to an enlargement and furrowing of the beds of valleys and to an increase in the steepness of their slopes. Retaining the uneven lengthwise profile of the valleys, inherited from the Ice Age of development of the relief, the ice has intensified this unevenness, with the result that in the lengthwise profile the depressions, many of which are occupied by lakes, alternate with ridges standing between them.

The breaking away and imbedding in the ice of rock fragments is due to shear planes that periodically arise as the ice moves. In Antarctica the rise of shear planes is closely bound up with the presence of moraines.

Shear planes evidently first rise on the border between the ice and the rock bed. Accumulating in the process of the movement of the sheet, the tension culminates from time to time in abrupt movements of the ice on the surface of the rock. Some of the ice melts during these movements. The appearance of water in the zone of contact between the ice and the rock leads to the rapid destruction of the latter.

As the ice climbs a protuberance of the rock bed, the shear planes that arise in the process may go beyond the ice-rock contact zone and join the ice in a magnitude that is determined by the rate of movement of the ice, the size of the ledge of the subglacial surface, etc. This enables the moraine constantly to move away from the ice-rock contact zone, enter the ice and form a layer of morain-containing ice in the base of the ice sheet. A feature of the structure of the moraine-containing ice is the alternation of layers of moraine-containing ice proper, «crystal» ice that is almost free of inclusions of air that arise during melting taking place during motion along the shear planes, milk-white ice with an abundance of air inclusions.

Inasmuch as the durability of the different ice varies, shear planes continue to form periodically in the contact zones between them, while within the layers the ice deforms in conformity with the law of the viscous-plastic flow. The optical axes of the crystals orient themselves perpendicularly to the plane of the displacement during deformation by the law of the viscous-plastic flow. Elongated air and mineral inclusions arrange themselves with their lengthwise axes in line with the movement of the ice. This explains the now well-known fact that in the moraine deposits the boulders all have their lengthwise axes oriented in the direction of the movement of the ice that had deposited the moraine.

In addition to the arrangement with the lengthwise axes in the direction of the movement of the ice, fragments of rock in the ice tend to acquire the specific form of a «flat iron», which is greatly facilitated by the large displacement along the shear planes within the moraine-containing ice.

At present all the moraine carried by the ice sheet of the Eastern Antarctic is borne to its edge and deposited either in the sea or along the periphery of the ice-free sections, forming terminal moraine ridges.

In the epoch of maximum glaciation, the moraine was carried beyond the boundaries of the present continent, forming terminal moraine ridges at a distance of 100 kilometres from the present-day edge of the ice sheet. Drigalski, Mill and Bowman islands are situated on the higher parts of the underwater ridges of terminal moraine. The surface of the ice sheet in the coastal zone was several hundred metres higher

and all the sections free of ice today were buried under ice. In the main, glaciation decreased probably through the sinking of the surface of the ice sheet and the cessation of the movement and melting of the ice in the more elevated sections of the sub-glacial surface. That this process of the liberation of the surface of the rock from the ice takes place is testified to by the thinness of the moraine deposits in them and the absence of terminal moraine formations, which would have taken shape if glaciation had decreased and ice-free sections had appeared as a result of the retreat of the edge of the ice sheet.

The moraine carried to the edge of the ice sheet at the present time is borne farther out into the sea by icebergs. A large quantity of moraine is deposited in the direct vicinity of the coast line, where there is an intensive breaking-up of icebergs transporting moraine in their underwater parts.

A considerable quantity of moraine is accumulated beneath the shelf glaciers, where the conditions possibly favour the melting out of moraine from the ice sliding seawards from the continent. The moraine dropped on the sea bed forms a special type of so-called iceberg deposits.

Terminal moraine ridges form where the end of the ice sheet climbs the ice-free surface. The reason for this is that moraine-containing layers emerge to the surface. Moraine accumulates as they melt and covers the ice core of the ridge formed by the usual moraine-bearing ice.

Along the edge of the ice-free sections of Eastern Antarctica there are two terminal moraine ridges, the older of which does not have an ice core and the younger, which is taking shape at the present, contains an ice core. The presence of these ridges of terminal moraine shows that a further decrease of the ice sheet has been taking place since the appearance of the ice-free sections. As a result of this the edges of the ice sheet containing large quantities of moraine broke away, the masses of moraine-containing ice that had broken away melted and moraine ridges were formed which had no ice core.

The moraine that is being accumulated at the present in the edge of the surface of the ice sheet is shaping a ridge with an ice core in its foundation.

MULTIPLE GLACIATION IN THE McMURDO SOUND REGION, ANTARCTICA A PROGRESS REPORT

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SUMMARY

At least four major Quaternary glaciations, each successively less extensive than the former, are recorded in the McMurdo Sound region of Antarctica. Deposits of the earliest recognized glacial advance occur high on ridges and flat areas. The deposits are 2,000 feet above the valley floor, are badly weathered, and have little or no morainal form. Ice of this glaciation filled all the valleys and must have filled McMurdo Sounn to an elevation of 2,000 feet.

Deposits of the next two succeeding glaciations are distributed around the Sound as well-preserved but considerably subdued moraines of both outlet and alpine glaciers. During the earlier of these two glaciations, alpine glaciers reached the expanded Koettlitz and Ferrar outlet glaciers. Outlet glaciers filled the southern part of McMurdo Sound to an elevation of about 1,000 feet. During the latter of these two advances many alpine glaciers did not reach the outlet glaciers.

The latest major glaciation is represented by well-preserved ice-cored moraines. Number and position of deltas in drained glacier-ice-blocked lakes suggest possibly three stillstands or minor advances during this glaciation. Radiocarbon dating of algae in drained ponds indicates a minimum age of 6,000 years for this glaciation.

NOUVELLE ESTIMATION DU VOLUME DE LA GLACE DE L'INDLANDSIS ANTARCTIQUE

A. BAUER

RÉSUMÉ

Les résultats des expéditions scientifiques de l'Année Géophysique Internationale permettent d'estimer le volume de la glace de l'indlandsis antarctique à 30 millions de km^3.

SUMMARY

The results of the scientific expeditions of the International Geophysical Year in Antarctica permit to estimate the volume of the antarctic ice cap to 30 millions km^3.

1. INTRODUCTION

Le volume de la glace de l'indlandsis antarctique est une donnée fondamentale de la glaciologie du globe. Bien des auteurs se sont chargés d'estimer ce volume avec les données disponibles. Citons les principaux : Bauer 1955, Antevs 1929, Cailleux 1952, Daly 1934, Dubois 1931, Flint 1947, Heinsheimer 1958, Meinardus 1926, Buinitsky 1953. Le volume estimé variait entre $7{,}8.10^6$ et 26.10^6 km^3 de glace.

Le développement extraordinaire des recherches entreprises sur l'indlandsis antarctique au cours de l'Année Géophysique Internationale a permis de préciser les conditions topographiques superficielles et les épaisseurs de la glace en de nombreux points. Ces données sont susceptibles de livrer une nouvelle estimation du volume de la glace de l'indlandsis antarctique.

2. MÉTHODE

En nous basant sur les résultats des sondages séismiques des Expéditions Polaires Françaises au Groenland, nous avons pu déterminer le volume de la glace de l'indlandsis groenlandais (Bauer 1954).

Les conditions altimétriques de la surface de cet indlandsis permettent de tracer la courbe hypsographique de la surface, d'où l'on déduit l'altitude moyenne de la surface. De même, l'altitude moyenne du socle rocheux est déduite de la courbe hypsographique du socle rocheux. La différence de ces deux altitudes moyennes donne l'épaisseur moyenne qui, multipliée par la surface de l'indlandsis, donne le volume de la glace couvrant le Groenland.

Cette méthode a été vérifiée par une intégration directe (Holtzscherer 1954). En effet, les sondages séismiques effectués pour déterminer l'épaisseur de l'indlandsis groenlandais ont été assez nombreux pour permettre de tracer quatorze coupes transversales le long des parallèles 63° à 80°N, de planimétrer leurs surfaces et d'en déduire le volume de glace par intégration.

Les résultats sont concordants : 2,6 à $2{,}7.10^6$ km^3 de glace, soit une différence de 5%.

Nous avons appliqué la première méthode à l'indlandsis antarctique (Bauer 1955). Avec les données dont nous disposions alors, nous avons pu tracer la courbe hypsographique de la surface et en déduire l'altitude moyenne.

Quant à l'altitude moyenne du socle rocheux à déduire de la courbe hypsographique de ce socle, comme les mesures d'épaisseurs étaient rares, nous avons adopté l'hypothèse de travail suivante : en première approximation, il était permis de supposer que les conditions altimétriques de la surface et du socle rocheux étaient semblables pour les indlandsis groenlandais et antarctique. Cette hypothèse permettait de tracer la courbe hypsographique du socle rocheux de l'antarctique déglacé et d'en déduire l'altitude moyenne. De ces deux altitudes moyennes on déduisait l'épaisseur moyenne et le volume de la glace de l'antarctique. Il était naturellement impossible de vérifier la valeur trouvée par une intégration directe. Nous allons utiliser la même méthode en utilisant les données nouvelles rapportées par les expéditions de l'Année Géophysique Internationale dans l'Antarctique.

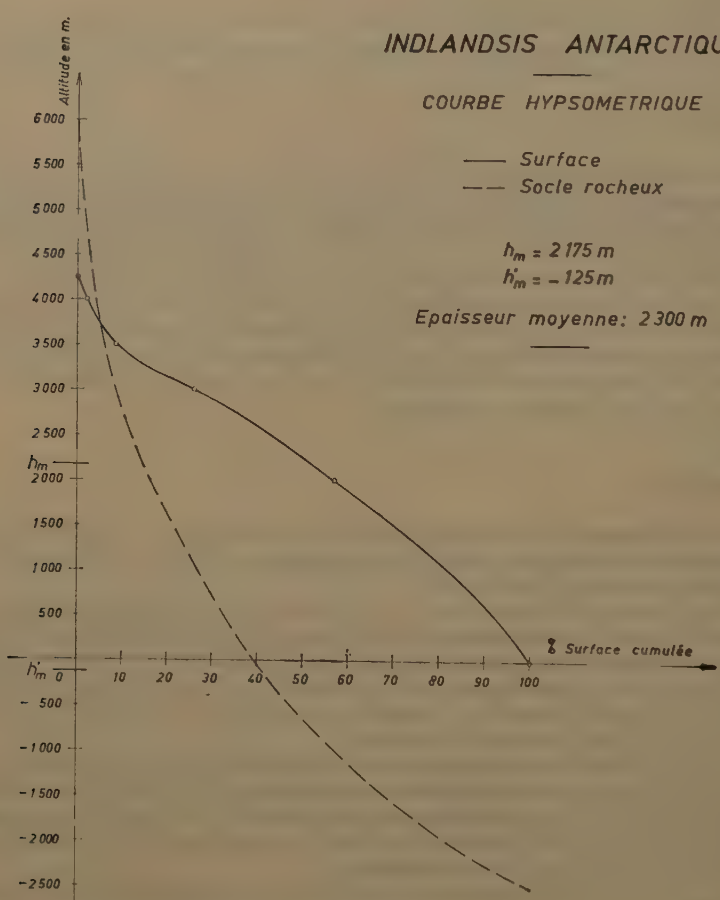

Fig. 1.

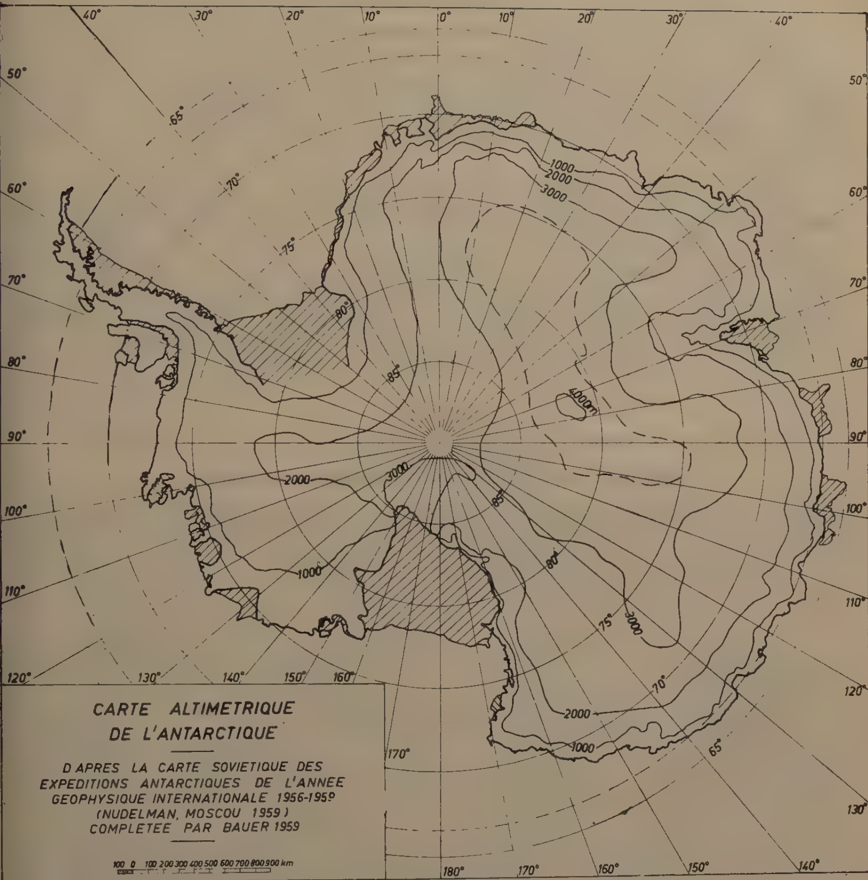

Fig. 2.

3. Données

Les résultats les plus spectaculaires des expéditions antarctiques de l'Année Géophysique Internationale ont été les suivants :

— culmination de l'indlandsis antarctique à 4 250 m près du Pôle d'Inaccessibilité, et non à 3 000 m comme on le supposait couramment;

— épaisseur de glace pouvant atteindre 4 000 m et socle rocheux pouvant se trouver à 2 500 m en-dessous du niveau de la mer.

Ces résultats sont consignés dans les rapports des expéditions dont nous ne citerons que les principales :

— expéditions américaines (Bentley & Ostenso, 1959)
— expédition du Commonwealth (Lister & Pratt, 1959)
— expéditions françaises (Imbert, 1958)
— expéditions soviétiques (Nudelman, 1959)

— expédition commune norvégienne, britannique et suédoise (Robin, 1958).

Les conditions altimétriques de surface ont été déduites de la carte de l'Antarctique au 1/10.000.000 des expéditions soviétiques (Nudelman, 1959). Les courbes de niveau ont été complétées par les données altimétriques des autres expéditions (Fig. 1). Même si notre tracé des courbes de niveau est quelquefois hasardeux, il ne peut changer beaucoup la courbe hypsographique de la surface que nous en avons déduite (Fig. 2).

Les ice-shelves font partie de l'indlandsis antarctique. La surface totale du continent antarctique de $13{,}4.10^6$ km² est à diminuer de $0{,}6.10^6$ km² des terres déglacéese: elle est donc de $12{,}8.10^6$ km².

Partant de l'hypothèse exposée précédemment, nous avons pu tracer la courbe hypsographique du socle rocheux et en déduire l'altitude moyenne. C'est la seule méthode utilisable. Il est bien évident que les sondages séismiques sont encore trop peu nombreux pour tracer des coupes et permettre une intégration directe.

4. Résultats

Conditions superficielles de l'indlandsis antarctique

Intervalles m	surface 10^4 km²	surface %	Surface cummulée %
0-1 000	269,3	21,0	100,0
1 000-2000	285,2	22,3	79,0
2000-3 000	393,5	30,8	56,7
3 000-3 500	224,5	17,5	25,9
3 500-4 000	104,8	8,2	8,4
4000-4250	2,1	0,2	0,2
	1 279,4	100,0	
soit	$12{,}8.10^6$ km²		

De ces valeurs, nous déduisons la courbe hypsographique (Fig. 2) et les valeurs suivantes :

Altitude moyenne de la surface		2 175 m
Altitude moyenne du socle rocheux		— 125 m
Epaisseur moyenne		2 300 m
Volume de glace	$12{,}8.2{,}3.10^6$ =	$29{,}5.10^6$ km³
Valeur en eau		$26{,}5.10^6$ km³
Lame d'eau répartie sur les mers du globe		73 m

Glaces de la Terre

	Surface 10^6 km²	Surface %	Volume glace 10^6 km³	Volume glace %	Volume eau 10^6 km³	Volume eau %	Lame d'eau m
Groenland	1,7	11	2,6	8	2,4	8	7
Antarctique	12,8	85	29,5	91	26,5	91	73
Autres glaciers	0,5	4	0,2	1	0,2	1	1
Total	15,0	100	32,3	100	29,1	100	81

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MOUVEMENTS GLACIOISOSTATIQUES DE L'ECORCE TERRESTRE

K.K. MARKOV

Résumé

Il est rationnel de porter attention sur deux preuves de la grande importance des mouvements glacioisostatiques de la surface de la Terre.

1. Les évaluations actuelles de l'altitude moyenne de l'Antarctide (2350 m) montrent encore plus clairement que les suppositions antérieures qu'elle dépasse de plus de 2 fois L/2 l'altitude moyenne des autres continents. Actuellement ces données peuvent être pour la première fois comparées à des données de recherches gravimétriques (S.A. Ouchakov et G.E. Lazarev), qui ont montré que l'Antarctide est isostatiquement équilibré. Conséquemment, il s'est produit un enfoncement du continent et un évasement des masses sous l'écorce amenés par la pression du manteau de glace dont la formation est justement la cause de la grande altitude de l'Antarctide.

2. Une comparaison des lignes côtières anciennes des bassins d'eau holocène de la Finlande avec les lignes côtières anciennes de la Mer Caspienne pléistocène montre, en tenant compte de l'échelle du temps, que les premières sont déformées plus que les secondes de 25 à 60 fois suivant la verticale.

Une aussi grande stabilité relative de l'écorce terrestre dans la région de la Caspienne comparativement avec la région du bouclier cristallin baltique peut être expliquée seulement par l'activation des mouvements verticaux de ce dernier à la suite de la décharge des glaces, c'est-à-dire par glacioisostasie.

Summary

It is rational to pay attention to two proofs of the great importance, played by glacioisostatic movements on the Earth's surface :

1. The present-day evaluations of the mean altitude of the Antarctica show more clearly than previous suppositions that the mean Antarctica altitude (2350 m) exceeds by more than two and a half times the mean altitude of all other continents. These data at present may be for the first time compared with data of gravimetric investigations (S.A. Oushakov and G.E. Lazarev), that revealed the fact that the Antarctica is in isostatic equilibrium. Therefore, here took place a sinking of the continent and an outflow of subcrustal masses, caused by pressure of the ice sheet, the formation of which was the reason of the Antarctica's great height.

2. A comparison of the ancient lines of the littoral in the holocene water basins of Finland with those of the Pleistocene Caspian sea reveels that the first ones are more deformed than the latter by 25-60 times along the vertical (taking into account the scale of time).

Such a great relative stability of the Earth's Crust in the region of the Caspian (in comparison to the region of the Baltic crystalline shield) can be explained only by activation of vertical movements of the latter by glacial discharge, i.e. glacioisostatically.

Le but du présent rapport est de souligner le rôle joué par le facteur glacioisostatique dans les mouvements tectoniques de la période quaternaire. Ce sujet a été déjà traité à plusieurs reprises, mais les données nouvelles recueillies ces derniers temps méritent d'être signalées à l'intension du lecteur.

1. Altitude moyenne de l'antarctide

Il y a quelque 50 ans (en 1909) W. Meinardus, géophysicien allemand, a constaté le premier que le niveau moyen de la surface glaciaire de l'Antarctide est nettement supérieur au niveau moyen de la surface des autres terres fermes. L'Antarctide est

située à une altitude trois fois supérieure à celle d'autres continents (altitude de 2000 ± 200 m au lieu de 700 m). Ces différences s'expliquent par une épaisse couche de glace reposant sur la surface des roches de fond en Antarctide comme au Groenland. Telle était l'hypothèse avancée par Meinardus. Cette ingénieuse hypothèse a été confirmée ces derniers temps, surtout au cours de la III[e] Année Géophysique Internationale, quand elle a pu enfin être étayée par des faits concrets. Je citerai quelques chiffres obtenus dernièrement et confirmant des déductions de Meinardus:

F. Makhatchek (1955) : Altitude moyenne de la terre ferme (Antarctide non comprise)	875 m
O.G. Sorokhtine, Y.I. Avsiouk et V.I. Koptev (1959) Altitude moyenne de l'Antarctide	3 000 m
Epaisseur moyenne de la couche de glace	2 200 m
Altitude moyenne des roches de fond au-dessous de la couche glaciale en Antarctide Orientale	800 m
K.K. Markov et G.G. Syrtsov.	
Altitude moyenne de l'Antarctide Orientale	2 500 m
Altitude moyenne de l'ensemble de l'Antarctide	2 350 m(*)

Il est certain que même ces données ne sont pas rigoureusement précises. Il existe des écarts entre elles. Cependant et malgré toutes les réserves qu'on pourrait faire, il est difficile de nier les faits suivants : 1) l'Antarctide est un continent exceptionnellement élevé et 2) l'altitude de l'Antarctide est due à la très grande épaisseur de sa couche glaciale. Ces données témoignent du rôle joué par le facteur glacio-isostatique dans les mouvements tectoniques contemporains.

On a procédé ensuite à l'étude de la répartition de la pesanteur en Antarctide. Les résultats obtenus (S.A. Ouchakov et G.E. Lazarev) peuvent être résumés comme suit.

La plus grande partie de l'Antarctide, à savoir l'Antarctide Orientale est une plate-forme au point de vue géologique. Elle peut être caractérisée à l'heure actuelle par un état proche de l'équilibre isostatique, existant également ponr les autres plates-formes continentales; cet équilibre s'est établi malgré la charge glaciale complémentaire. Cependant la compensation n'a pu avoir lieu qu'au dépens de l'abaissement de la surface des roches de fond et du refoulement de la substance subcorticale. Donc, on peut déduire de l'état d'équilibre isostatique ou voisin de ce dernier que l'Antarctide s'est abaissé sous le poids de la glace.

L'épaisseur moyenne de la glace en Antarctide pouvant être estimée à 2 200-2 400 m et le rapport des densités des roches et des glaces étant de 3 : 1, on peut évaluer l'abaissement glacio-isostatique de l'Antarctide à 750 m en moyenne. L'altitude de l'Antarctide s'est accrue par accumulation d'une couche glaciale «légère», trois fois plus épaisse que l'abaissement du fond «lourd» du continent.

Les savants ayant effectué des recherches sur le Groenland du nord sont arrivés aux mêmes conclusions; là aussi on a constaté l'équilibre isostatique de l'écorce terrestre plus la couche de glace, tandis que le fond de l'île a une forme eu calice. On sait que le Groenland rappelle l'Antarctide par l'altitude moyenne de sa surface (2 130 m) et par l'épaisseur de la couche de glace (1 500 m). L'Antarctide s'est abaissée sous le poids de la glace en ayant conservé cependant (l'Antarctide Orientale) une altitude moyenne de la surface du fond rocheux proche de l'altitude moyenne des autres continents (800 et 875 m respectivement). Avant la glaciation son altitude (800 + 750 = 1 500 m environ) était deux fois plus élevée que le niveau des autres continents. Il est probable que les phénomènes glacio-tectoniques n'ont pas été les seuls à déterminer l'altitude de la surface des roches de l'Antarctide, mais que celle-ci est dûe à

(*) Pour les détails voir l'article de l'auteur cité dans la bibliographie.

d'autres processus également dont l'analyse sort du cadre du présent rapport. Compte tenu de cette remamrque on peut affirmer que la gravimétrie de l'Antarctide (et du Groenland) correspond à l'abaissement glacio-isostatique des roches de fond de grande terre.

2. Vitesse de soulèvement de la Fennoskandia

Citons quelques chiffres caractéristiques de la vitesse des mouvements verticaux contemporains de l'écore terrestre. En Finlande même la vitesse actuelle, déjà ralentie, du mouvement vertical de l'écorce terrestre est plus grande que pour la majorité des localités de la plate-forme russe. Exprimée en millimètres par an, cette vitesse est égales à :

Finlande		Plate-forme russe I(**)		Plate-forme russe II(***)	
Vaasa	9,0	Léningrad	— 0,1	Vilnius	+ 3,8
Tornio	8,4	Volkhov	— 4,3	Minsk	— 1,6
Oulu	8,2	Dno	— 1,8	Kiev	+ 10,4
Khopio	5,8	Bologīé	— 3,6	Stalingrad	+ 1,3
Turku	5,2	Moscou	— 3,9	Odessa	— 5,1
Enare	4,9	Orel	0,0	Sébastopol	– 0,5
Helsinki	2,95	Koursk	+ 3,6	Kertch	+ 0,2
		Kharkov	+ 3,6	Taganrog	+ 0,7

En outre les mouvements de l'écorce terrestre en dehors des limites du bouclier cristallin Baltique ne suivent pas un plan aussi simple que celui qui attire toujours l'attention à l'intérieur de ce bouclier. L'explication la plus probante de cette différence a été donnée par les auteurs de l'ouvrage sur les mouvements verticaux contemporains de l'écorce terrestre, avec lesquels je suis entièrement d'accord : «Il est probable que lors de la désagrégation de la vaste zone de soulèvement de nature isostatique, il s'est produit une différenciation des mouvements tectoniques suivant les éléments structureotectoniques» (p. 86).

Ainsi les vitesses contemporaines du mouvement de la surface de la Finlande sont dans l'ensemble plus grandes que celles de la surface de la plate-forme russe et contrairement à ces dernières ont un signe constant.

Cependant, il ne faut pas oublier qu'il y a 8 000 ans la surface de la Finlande s'élevait de 6 à 7 fois plus rapidement qu'à l'heure actuelle. Nous ne connaissons pas quelle a été à cette époque, proche de la période glaciale, la vitesse de soulèvement de la plaine russe, sauf peut-être pour les côtes orientales de la Baltique.

En tout cas, si la vitesse élevée de soulèvement de la surface constatée il y a 8 000 ans pour la Finlande peut être expliquée par la disparition des masses de glace, on ne peut supposer d'accélération de même nature pour les régions de la plaine russe, non recouvertes par les glaces, et pour lesquelles la vitesse de soulèvement contemporaine est cependant faible (voir plus haut).

La confrontation des mouvements des côtes des mers Baltique et Caspienne est encore plus intéressante. Le mouvement de l'écorce terrestre dans la région du bouclier

(**) Suivant E. Keeriejnen.
(***) Suivant l'ouvrage : «Mouvements verticaux contemporains de l'écorce terrestre».

cristallin Baltique, ainsi que de la mer Caspienne, peut être évalué par le caractère plus ou moins accidenté des cotes anciennes de cette dernière ou plutôt par la dénivellation de certains points par rapport aux autres.

P. F. Fédorov estime que les niveaux des terrasses des terrasses de Khvalynsk et d'autres plus récentes encore de la Caspienne varient extrèmement peu sur toute l'étendue de la côte orientale de cette mer. Dans son ouvrage de 1957, il cite les données sur la dénivellation de certains points faisant partie de la même ligne côtière et de massifs de dépôts correspondants.

Epoques			
de Bakou	620 m	début de l'époque de Khvalynsk	15 à 20 m
de Khazar	130 m	fin de l'époque de Khvalynsk	11 m

Les superficies de la mer Caspienne et de la Finlande sont à peu près égales, ce qui facilite la comparaison des données.

L'élément suivant important est l'âge des lignes côtières. La ligne côtière ancienne remontant à 8 000 ans est la ligne Ancile tandis que celle comptant 5 000 ans est la ligne Lithorine. En ce qui concerne la mer Caspienne, toutes les lignes côtières énumérées ci-dessus sont nettement plus anciennes que les lignes côtières de la Baltique dont nous venons de parler. Supposons que l'âge du bassin de Khvalynsk, à la fin de cette époque correspond à l'époque de glaciation de Voldaï, à sa seconde moitié pour être plus exact. Situons le bassin de Bakou dans le bas quaternaire. Nous aurons alors les valeurs des déformations et les âges suivants;

Bassin de Khvalinsk (à la fin de cette époque)	11 m	20 000 ans
Bassin de Bakou	620 m	500 000 ans
Mer Lithorine	100 m	5 000 ans
Lac Ancile	280 m	8 000 ans

Si nous comparons les lignes côtières de la mer Ancile et du bassin de Khvalynsk nous constatons que, malgré que l'âge de la seconde et la durée de sa déformation sont de deux et demi fois supérieurs aux caractéristiques correspondantes de la première, la côte Ancile est déformée 25 fois plus que la côte de Khvalynsk. La ligne côtière de Bakou est déformée trois fois plus que l'Ancile, tandis que son âge est de 60 fois supérieur.

Il ne faut pas oublier non plus que la dépression de la mer Caspienne est située dans son ensemble dans une partie plus mobile de l'écorce terrestre que le bouclier cristallin Baltique.

Les écarts dans les vitesses des mouvements verticaux de l'écorce terrestre du bouclier Baltique et de la région Caspienne ont été constatés non seulement par l'auteur mais également par P.F. Fédorov et Y.A. Mechtchériakov. Ce dernier écrivait que le bouclier Baltique s'est élevé au cours des 12 000 à 15 000 ans d'autant que le Caucase en 750 000 ans. Il est difficile d'expliquer cette différence par des mouvements oscillatoires d'un rayon aussi grand qu'ils ne se soient que faiblement manifestés dans la région Caspienne. Le soulèvement de la voûte du bouclier Baltique fait partie également des mouvements à grand rayon. Il est plus naturel de supposer que le soulèvement de ce bouclier a eu lieu dans la période postglaciale à une vitesse aussi grande parce qu'il à été activé par la disparition de la couche glaciaire, c'est-à-dire que ce soulèvement a été (et continue d'être) un soulèvement essentiellement glacio-isostatique.

Ainsi deux constattaions faites au cours des dernières années, confirment une fois de plus l'importance des mouvements glacio-isostatiques de l'écorce terrestre. Ce sont : 1) la grande altitude des régions glaciaires contemporaines et 2) la grande vitesse de leur soulèvement.

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STUDIES IN GLACIAL METEOROLOGY AT LITTLE AMERICA V, ANTARCTICA

I. Net radiation, heat balance, and accumulation during the winter night 1957

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ABSTRACT

The net long-wave radiation at Little America V is discussed for the winter night, April 21 to August 20, 1957. Under clear skies the outgoing net radiation decreases linearily with decreasing temperature. A quarter of the hourly mean values, especially those with a low overcast, show incoming net radiation. The average radiative energy loss equals — 23 ly/day. Based on daily measurements of snow temperature to a depth of 8 meters, the average heat loss of the uppermost 8 m of the snow was found to be only about — 6 ly/day. Therefore about + 17 ly/day must have been supplied by convective processes from the atmosphere. Using LILJEQUIST's method, it was found that + 12 ly/day were supplied in the form of sensible heat. The heat budget of the snow can be outbalanced by sublimation heat, released in the process of sublimation of 1.02 g of hoar frost per square centimetre during the winter night, which seems to be a reasonable estimate.

The accumulation, erosion and the stratigraphy of the snow was studied in detail in connection with weather events. The accumulation takes place mainly in spring, summer and fall, whereas during winter erosion predominates. During the one-year's period February 1957 through January 1958 the net accumulation amounted to 58 cm of snow with an average density of 0.345 g cm^{-3}, and a water equivalent of 20 cm. This figure agreed fairly well with the measured precipitation of 22.5 cm water equivalent.

ZUSAMMENFFASSUNG

Die langwellige Strahlungsbilanz in Little America V wird für die Winternacht 21. April bis 20. August 1957 diskutiert. Bei klarem Himmel nimmt die negative Strahlungsbilanz mit abnehmender Temperatur linear ab. Ein Viertel der stündlichen Mittelwerte, besonders die mit tiefer Bewölkung, zeigen langwellige Einstrahlung. Der mittlere Energieverlust durch langwellige Ausstrahlung beträgt — 23 ly/Tag. Basierend auf täglichen Messungen der Schneetemperatur bis zu einer Tiefe von 8 m wird ein mittlerer Wärmeverlust der obersten 8 m mächtigen Schneeschicht von etwa — 6 ly/Tag gefunden. Es müssen daher etwa + 17 ly/Tag durch konvektive Prozesse aus der Atmosphäre zugeführt worden sein. Mit der von LILJEQUIST angegebenen Methode wurde berechnet, daß etwa + 12 ly/Tag in Form von fühlbarer Wärme zugeführt worden sind. Die Wärmebilanz des Schnees wird durch Sublimationswärme ausgeglichen, die bei der Netto-Sublimation von 1.02 g Reif pro Quadratzentimeter während der Winternacht frei wird. Das scheint eine vernünftige Größenordnung zu sein.

Die Akkumulation, Erosion und die Stratigraphie des Schnees wurde im Zusammenhang mit den Witterungsereignissen detailiert studiert. Die Akkumulation erfolgt vorwiegend im Frühjähr, Sommer und Herbst, während im Winter Erosion vorherscht. Wàhrend der einjahrigen Periode Februar 1957 bis Januar 1958 betrug die Netto-Akkumulation 58 cm Schnee, mit einer mittleren Dichte von 0.345 g cm^{-3}, und einem Wassergehalt von 20 cm. Diese Zahl stimmt mit dem gemessenen Niederschlag von 22.5 cm Wasseräquivalent recht befriedigend überein.

1. Introductory remarks

During the US-IGY-Antarctic Expedition 1957-58 material for the heat balance and mass balance of Antarctic snow fields was collected between February 1, 1957, and February 3, 1958, mainly at Little America *V* (78°11'S, 162°10'W, 44 meters elevation above sea level). Measurements were performed and records were made of the following quantities: normal incidence solar radiation, using an actinometer Linke-Feussner; global radiation and albedo, with solarimeters Moll-Gorczynski, and long-wave radiation, using a net radiometer Schulze, both thermopiles of which were recorded separately. A Leeds & Northrup Speedomax Type G 6-point recorder was kindly placed at my disposal by the US-Weather Bureau. Calibrations of the thermopiles were made whenever the weather-situation was favourable. The thermopiles were found to become more sensitive with lower temperatures as well as with lower solar altitudes. When solar altitudes exceeded approximately 16 degrees the calibration factors for clear sky became almost identical with the factors for isotropic radiation. When exposed over a dry snow surface the downfacing thermopiles continuously receive isotropic radiation, whereas the upfacing thermopiles receive this type of radiation only under conditions of a dense uniform overcast. Therefore, different calibration factors must be applied for the upfacing than for the downfacing thermopiles with a clear or broken sky, as long as the solar altitude does not exceed 16 degrees. Because of these facts, which were not always taken into account in the past, it is, in principle, impossible to record the albedo or the net radiation directly in the polar regions, where the sun remains at low angles for a long time. Both quantities can only be calculated from separate records of up- and downfacing thermopiles, using different calibration factors, depending on solar altitude, cloudiness, and temperature. Considerable difficulties arise from the formation of hoar-frost or rime on the glass- or polyethylene-bulbs covering the thermopiles. Only checking and cleaning the bulbs as often as possible enables one to correct for the error introduced by a frost-cover. With bright sunshine a frost-cover normally tends to increase the recorded radiation; only when heavy rime forms, the radiation is decreased, as it is always the case with diffuse radiation. More details about calibration of and experiences with the radiation instruments will be published in another paper.

Some preliminary results of these radiation studies, dealing with measurements of normal incidence radiation and albedo, have been published already (H. Hoinkes, 1960B). The results of the global radiation and albedo, as recorded at Little America *V*, will be published in full detail elsewhere. Since the calculation of the net radiation for the summer-period is still in progress, the present paper shall deal with the net long-wave radiation during the period when the sun was continuously below the horizon, between April 21 and August 20, 1957, referred to in this paper as the *winter night*. For the same period an attempt is discussed to estimate the heat balance of the snow surface near Little America *V* on the Ross Ice Shelf. For this purpose use is made of temperature data recorded by the US-Weather Bureau with «thermohms» at six different levels, and of thermohm-readings of snow temperatures in six different depths, down to eight meters, performed by A.P. Crary and R.L. Chappell, respectively.

2. The net long-wave radiation during the winter night

For the winter night the records for both thermopiles and the temperature of the net radiometer Schulze were evaluated in mean ordinates for half-hourly intervals, using Local Apparent Time (LAT). Whenever the recorded trace was influenced by

frost, corrections had to be applied, the amount of which was known from frequently checking and cleaning the polyethylene-bulbs. The ordinates were multiplied with the respective calibration factors for isotropic radiation, depending on the instrument's temperature. These values were added in pairs and the algebraic sums of the incoming radiation from the upper and the outgoing radiation from the lower hemisphere computed, thus giving mean hourly values of the net long-wave radiation in Langleys (Langley = cal/cm^2). The calibration factors for both thermopiles of the net radiometer were established by numerous comparative readings with the Linke-Feussner actinometer on clear days, the net radiometer being shaded from the sun by a small plate. The results are given in units of the International Pyrheliometric Scale 1956 (Radiation Commission of the IAMAP, 1958). In a previous report given at the Radiation Symposium of the IAMAP, Oxford, July 1959 (H. Hoinkes 1960 A), calibration factors were used as estimated with a black body source of radiation, which turned out to be too low by about 20 percent. A new and better calibration method with a ventilated black body source showed satisfactory agreement between both calibration methods, i.e. for long-wave and for short-wave radiation. Therefore, the figures for the net long-wave radiation given in part E of the above mentioned report (H. Hoinkes 1960 A) should be replaced by the results of the present paper. When comparing the net long-wave radiation given in this report with the values of other authors, it should be noted that the values herein presented are expressed as ly/min (obtained by dividing the mean hourly values by 60), whereas net radiation values for polar regions previously published were instantaneous values.

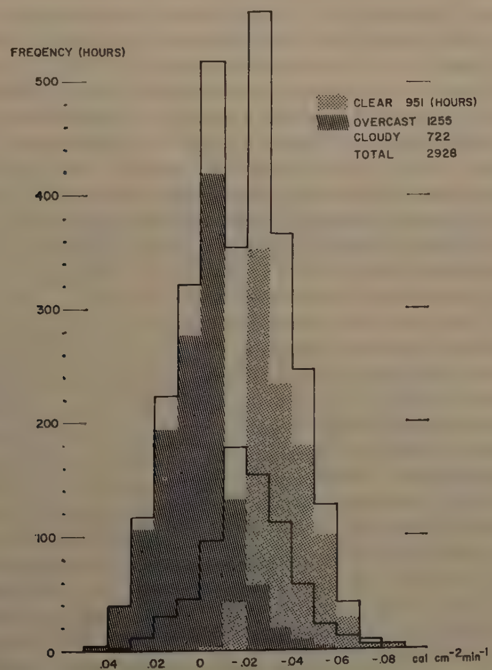

Fig. 1 — Frequency distribution of hourly mean values of net long-wave radiation for clear, cloudy and overcast sky conditions. Winter night, April 21 to August 20. 1957, Little America V, Antarctica.

Table 1 gives the frequency distribution for the 2928 hourly intervals of net radiation between April 21, 00 hours and August 20, 24 hours LAT in ly/min. Two peaks stand out quite clearly, the mean value of − 0.016 ly/min falling in the interval between. 2224 hourly intervals or 76 per cent of the four winter months have outgoing net radiation, and 704 hourly intervals or 24 per cent incoming one, the relation varying between 84 per cent outgoing and 16 per cent incoming in May and 72 per cent outgoing and 28 per cent incoming in July. The frequency distribution published by G.H. LILJEQUIST (1956) for Maudheim showed only 9.3 and 14 per cent of incoming net radiation for the winter half-years of 1950 and 1951 respectively, but as the author mentioned, more observations were made in clear weather. The frequency distribution will vary from year to year, depending on the different cloud conditions and the frequency and intensity of inversions.

As shown in fig. 1, the two peaks of the frequency distribution belong to overcast and clear sky-conditions, respectively. Observations of the cloud cover were carried out at full hours GCT, but sometimes they might not be representative of the whole hourly interval for which the net radiation has been computed. With cloudless sky the frequency maximum of the net radiation falls in the interval − 0.02 to − 0.03 ly/min, and the same holds for a sky cover of one tenth, the mean of 951 hourly intervals with clear sky being − 0.0354 ly/min, because of the skewness of the frequency distribution. With an overcast of low clouds the maximum frequency of the net radiation is found between zero and − 0.01 ly/min, the mean of 686 hourly intervals being + 0.0045 ly/min. It is particularly interesting to see the very pronounced frequency maximum between zero and − 0.01 ly/min for the 246 hourly intervals with heavy blowing snow (Table 1), so that neither the amount nor the kind of clouds could be observed. The average of − 0.0013 ly/min shows a slight energy loss to be typical of non-inversion conditions, even with the air densely filled with drifting snow particles. The average for 323 hourly intervals with an overcast of middle and high clouds is slightly negative as well, but the average for the 1255 hourly intervals with an overcast of any kind remains slightly positive (+ 0.0018 ly/min).

There is a definite relation known to exist in polar regions between the net long-wave radiation and the temperature at instrument level under cloudless skies. Fig. 2 contains 824 mean hourly values of outgoing net radiation, entered in intervals of 4°F of the recorded temperature of the net radiometer. Not the single points are plotted, but for the sake of clarity only the numbers of points in distinct squares. As previous investigators found, there is a high scatter in the single points; nevertheless the group-means, computed for intervals of 10°F (see Table 2) show a linear relation, the net radiation decreasing with decreasing temperature. This fact can be observed in each single period of consecutive clear days. Our values are nearly the same as those given by H. WEXLER (1941) for Fairbanks (66°N, 148°W), whereas the values measured by G.H. LILJEQUIST (1956) at Maudheim (71°S, 11°W) are considerably higher. Actually one should expect a much closer agreement between Maudheim and Little America *V*, because both stations were situated on Antarctic Ice Shelves. A thorough study of the net radiation in connection with the radiosonde-data from Little America *V* is in progress, which possibly may throw some light on this difference.

Table 3 contains daily totals of the net radiation and daily means of cloudiness, calculated from hourly observations. The highest daily energy loss from the snow surface due to net long-wave radiation amounts to − 77 ly, the highest energy gain to + 30 ly. On 20 days or 16 percent of the period under consideration the snow surface receives energy from the atmosphere, and on 102 days (84 per cent) an energy loss takes place. The highest monthly total net long-wave radiation amounts to − 854 ly in May, the lowest to − 609 ly in June. During the 122 days from April 21 to August 20, 1957, the net long-wave radiation amounted to − 2809 ly, the average radiative energy loss during the winter night equals − 23 ly/day, which is only half

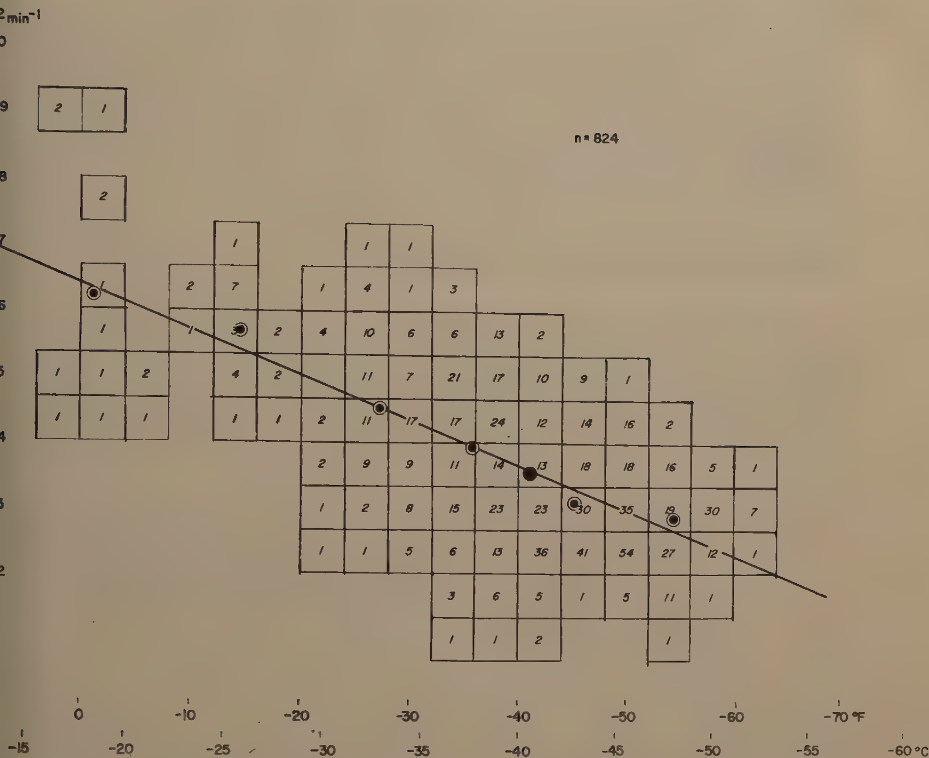

Fig. 2 — 824 hourly mean values of outgoing net radiation with cloudless sky as a function of the temperature of the net radiometer. Figures indicate the number of points in each square, group means are marked as dots. Winter night, April 21 to August 20, 1957, Little America V, Antarctica.

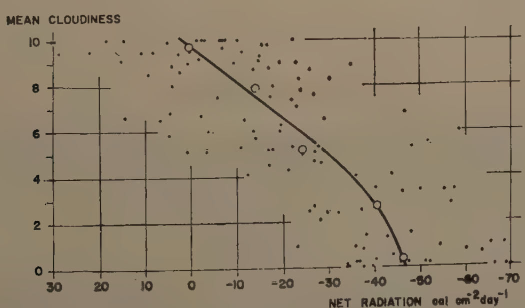

Fig. 3 — Daily totals of net long-wave radiation as a function of daily average cloudiness. Dots indicate group means. Winter night, April 21 to August 20, 1957, Little America V, Antarctica.

the amount adopted by G. H. LILJEQUIST (1956) for the winter months April to August at Maudheim. Fig. 3 shows the daily totals of net long-wave radiation as a function of daily average cloudiness. Some sort of a non-linear relation seems to exist, as indicated by the group means, the negative daily totals of the net long-wave radiation decreasing with increasing daily average cloudiness. The scatter of the single points is too great to permit further refinement of the curve shown in fig. 3. For example, a daily total for the net long-wave radiation between —20 and —25 langleys may be observed at any average cloudiness.

3. The heat balance during the winter night

In view of the fact that no melting took place during the entire winter night, the only change caused by the heat balance was a change in snow temperature. On April 5, 1957, six resistance thermometers were placed into the snow at different depths by A. P. CRARY. The temperature was measured once a day, using a Wheatstone

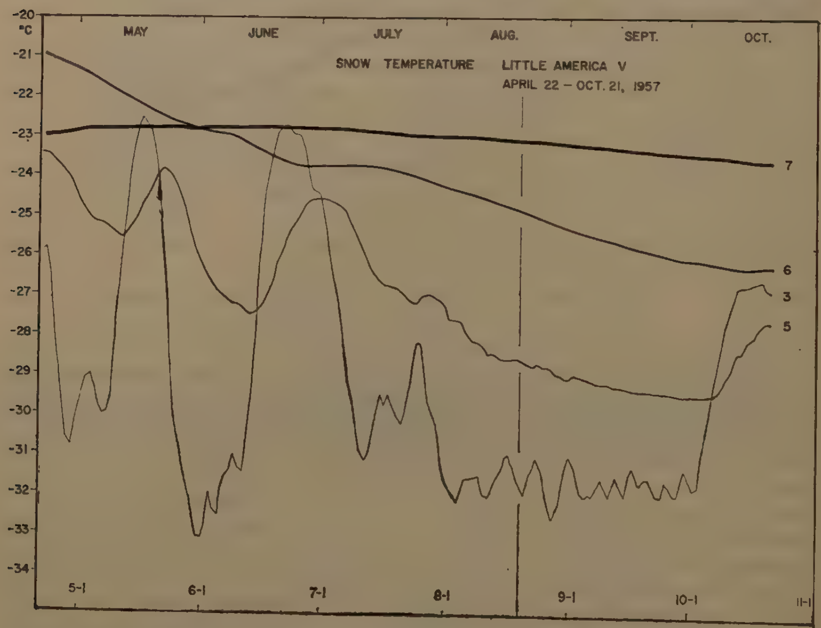

Fig. 4 — Snow temperature (°C), as measured with thermohms at Little America V, Antarctica. Between April 22 and Oct. 21, 1957, the depth changed from 0.65 m to 1.13 m for thermohm no. 3, from 1.60 m to 2.09 m for no. 5, from 4.09 m to 4.57 m for no. 6, and from 7.74 m to 8.22 m for no. 8.

Bridge Temperature Indicator Leeds & Northrup, until October 21, 1957, from which date on the Wheatstone Bridge was used by the departing Ross Ice Shelf Traverse-party. The recording thermohm no. 1 (of the US-Weather Bureau) was kept at the surface and could be read only during the winter night. On April 21, 1957, the depths of the thermohms were as follows: no. 2 at 0.10 m, no. 3 at 0.65 m, no. 4 at 0.85 m, no. 5 at 1.60 m, no. 6 at 4.09 m and no. 7 at 7.74 m. Due to accumulation of snow at the measuring site, the depths increased by an average of 26 cm between April

21 and August 20, and by an additional 22 cm between August 20 and October 21, 1957. Fig. 4 shows the snow temperatures, as measured with thermohms no. 3, 5, 6 and 7 for the period April 22 to October 21, 1957, in degrees Centigrade. The most remarkable features are two period of strong warming, one occurring during the first half of May, the other during the first two thirds of June. The warmings were caused by advection of warmer air during strong cyclonic activity and could be traced down to a depth of almost four meters. The deepest thermohm at an average depth of eight meters reached its highest temperature of $-22.8°C$ at about mid-winter-time, and fell to $-23.1°C$ on August 20, and to $-23.6°C$ on October 21, 1957.

The negligibly small temperature change at eight meters depth permits the estimation of the net change in heat content of the snow, according to the expression

$$\Delta Q = c \cdot \varrho \int_0^{\infty} \Delta T_{(z)}\, dz \quad \text{ly}$$

In order to calculate the average change of temperature for the uppermost 800 centimeters of snow cover, the temperatures were plotted as a function of depth between

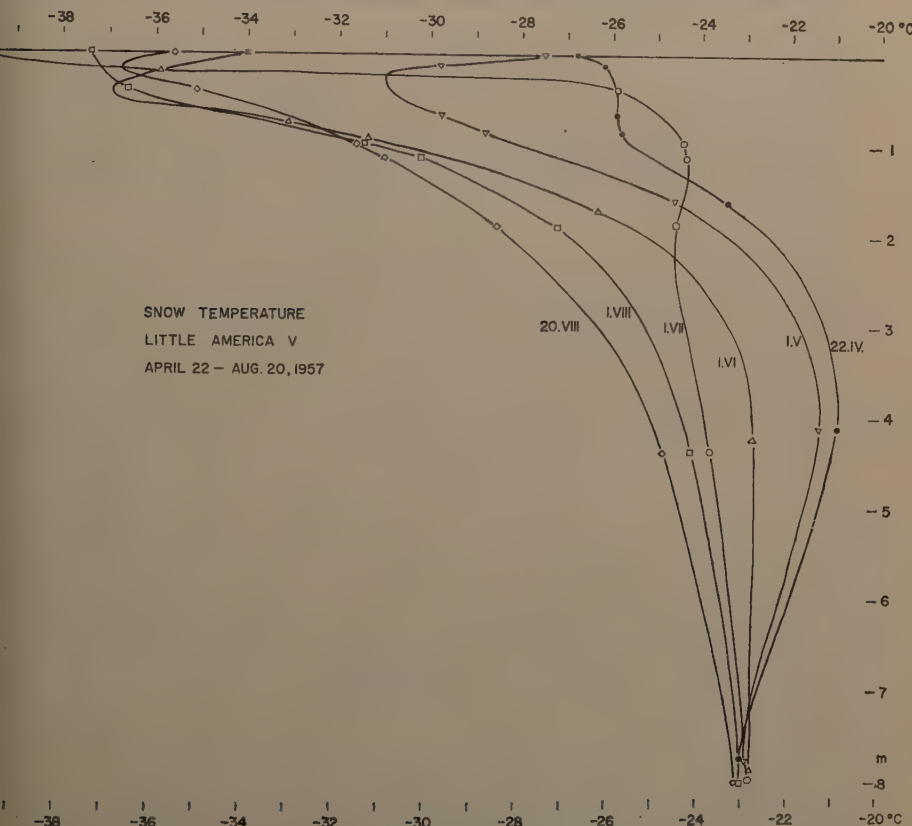

Fig. 5 — Snow temperature (°C) at Little America V, Antarctica, as a function of depth, on April 22, May 1, June 1, July 1, August 1, and August 20, 1957.

April 21 and August 20, 1957 (Fig. 5). During the four months of the winter night the average cooling amounted to -4.1°C, or about 1°C per month, despite an average warming during June of $+0.8$°C (compare Table 4). Using mean values for the specific heat of the ice and for the density of the uppermost 800 cm of snow, i. e. $c = 0.46$ cal.g^{-1}. degC^{-1} and $\varrho = 0.45$ g.cm^{-3} respectively, the net energy loss from the snow is calculated to be about -679 ly during the winter night. Because the net long-wave radiation during the same period caused an energy loss of -2809 ly, an energy amount of $+2130$ ly had to be supplied by convective transfer from the atmosphere, either in the form of sensible or of latent heat.

In order to verify the net energy gain, caused by convective processes, the flux of sensible heat due to turbulent mixing had to be calculated, using the expression

$$Q_s = c_p A \frac{\partial \vartheta}{\partial z} t \quad \text{ly},$$

c_p being the specific heat of dry air at constant pressure, A the «Austausch»- coefficient, ϑ the potential temperature, z the height above the snow surface, and t the time in seconds. The «Austausch»- coefficient for logarithmic wind profiles is given by the formula $A = \varrho\, 1 u_*$ g.cm^{-1} sec^{-1}, with ϱ the density of the air, 1 the mixing lenght, and the friction velocity

$$u_* = \frac{u_z}{(\log z - \log z_0)\, 5.75} \quad \text{cm.sec}^{-1},$$

z_0 being the roughness parameter. Very detailed studies of wind- and temperature profiles were performed by Paul C. Dalrymple (1960 at Little America V in 1957 and at the South Pole in 1958. Only a few preliminary results of these studies were

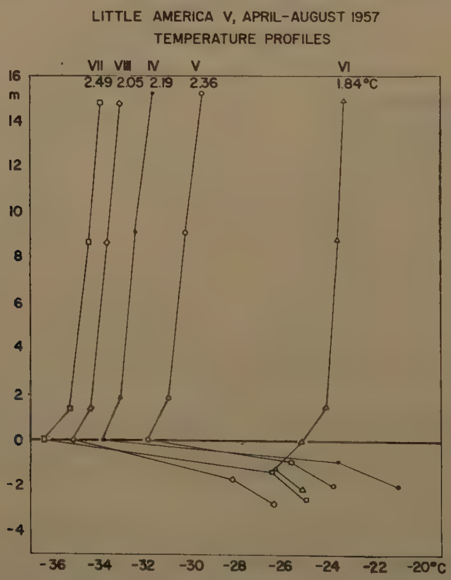

Fig. 6 — Profiles of monthly mean temperature from minus 2 to plus 15 meters, computed from hourly readings of thermohm data, recorded by the U.S. Weather Bureau at Little America V, Antarctica, Figures on top of the profiles indicate mean temperature differences between surface and 15 m height for the months April through August, 1957.

known to the present author at the time the following calculations of the heat balance were made.

A complete set of temperature observations was available for the time in consideration from six thermohms, set up by the US-Weather Bureau, the values of which were recorded continuously in the form of 13 profiles for every hourly interval. Three ventilated thermohms were placed on a mast at average heights of 15, 9 and 1.6 meters respectively, one was kept on the surface and the two remaining ones were placed at a depth of one and two meters below the surface; the depth of the latter two changed due to accumulation of snow to 1.7 and 2.7 meters as the winter proceeded. The records were evaluated in hourly values and worked up statistically; the full results of this study shall be discussed elsewhere. Average monthly temperature profiles from minus two to plus fifteen meters are given in fig. 6. They show, as one would expect, the lowest temperature at the snow surface, thus indicating an average heat flux towards the surface from within the snow by heat conduction, and from the air mainly through turbulent mixing. The average temperature difference between the surface and a height of fifteen meters for the months April through August is 2.19°C, with the highest average monthly value 2.49°C in July and the lowest 1.84°C in June.

The speed and direction of the wind was recorded continuously with an Aerovane, placed nine meters above the ground. Using average monthly wind velocities, and assuming a logarithmic profile with the roughness parameter $z_0 \simeq 0.01$ cm, one obtains average values for the friction velocities and for the «Austausch»-coefficient, the latter being of the order of magnitude $A \simeq 13$ g.cm^{-1}sec^{-1} at nine meters height. Using average temperature gradients, as derived from the above mentioned profiles for a height of nine meters, it is possible to calculate the heat flux from the air to the snow due to turbulent mixing. The resulting energy quantities are about ten times too high, but they will be reduced somewhat when corrections for the stable stratification of the air are applied. The failure of this simple method is partly due to the fact that the temperature profiles are not logarithmic ones, but rather show a tendency towards linearity. This is clearly seen in the average monthly profiles (fig. 6), and comes out even better in single profiles. The situation becomes more complicated by the fact that the shape of the temperature profile is also a function of wind direction. Especially with the prevailing winds from the interior of the continent, the temperature profile is frequently a linear one, and the same tendency is to be seen in the wind profiles. Only after the results of the detailed studies of temperature- and wind- profiles (P. C. Dalrymple, 1960) become available, can a new attempt be made to calculate the turbulent heat flux from temperature- and wind-profiles.

G.H. Liljequist (1957 A) experienced similar difficulties in calculating the turbulent transfer of heat at Maudheim. He arrived at the conclusion, that «the turbulent flux of heat in well developed inversions is proportional to the wind speed at a reference level (10 m), but independent of the temperature gradient». It is of high interest to check Liljequist's formula for the turbulent heat flux $Q = 0.0058.u_{10}$ ly.min^{-1} (where u_{10} is the wind speed at the 10 m-level in meters per second) with the observations from Little America V, because it would be very helpful for future work to have easier means for estimating the energy balance in polar regions. Liljequist (1957 B) has outlined his method for actual weather conditions. Following this procedure, we consider as a «well-developed inversion» a temperature-difference between 15 m and 1.6 m greater than 1.5°C. In these cases the average hourly wind velocity in meters per second is counted full. For weak inversions, i.e. temperature differences 15 m minus 1.6 m between o.5°C and 1.5°C, the hourly wind velocity is halved. The wind velocity for non-inversion conditions, i.e. temperature differences smaller than 0.5°C, does not count. These wind velocities were added for monthly periods and multiplied by 60.C, with C = 0.006 for an anemometer height

of nine meters, thus giving the amount of heat supplied by turbulent mixing in langleys during the period in question. For the period April 21 to August 20, 1957, an amount of +1440 ly was supplied in the form of sensible heat.

In order to outbalance the heat budget of the snow, we still need an amount of +690 ly, which most probably was supplied by latent heat, released in the process of sublimation of 1.02 grams of hoar frost per square centimeter during the four months of the winter night. This seems to be the only means for estimating evaporation from the snow or sublimation of hoar frost, since it is still too difficult to measure gradients of the specific humidity with an accuracy sufficient to calculate the turbulent humidity transport under the low temperatures prevailing during the winter night. The amount of 1.02 grams of hoarfrost per cm^2 is a minimum estimate, since surely not all of the heat released in the process of sublimation is used for heating the snow. It is of the same order of magnitude, only slightly less, than the sublimation calculated by V. SCHYTT (1960) from the snow temperatures down to a depth of 700 cm at Maudheim. This result indicates that LILJEQUIST's method to compute the turbulent heat flux gives values of the right order of magnitude. Even though the physical process of turbulent heat transfer under varying inversion conditions is extremely complicated and not yet completely understood, LILJEQUIST's method apparently can be used, wherever quantitative estimates are sufficient.

Table 4 contains the estimated heat balance of the uppermost 800 cm of snow near Little America V for the periods April 21-30, May, June, July and August 1-20, covering the 122 days between the last sunset and the first sunrise. Because of the insignificance of the heat flow at 800 cm, the balance is about representative of the Ross Ice Shelf in the vicinity of Little America V, assuming no variations at the bottom. It is seen that sublimation of hoar frost takes place mainly in May and June, whereas evaporation and sublimation apparently compensate each other more or less during the remaining two months. The heating of the snow during the two periods in May and June (see fig. 4) is therefore at least to a certain degree caused by released sublimation heat. The amount of hoar frost could not be measured, but whenever the radiation instruments were checked (on the average of three times a day) a note was made about the frost cover on the bulbs. From Table 5 it is seen that during May and June the number of cases «with frost on the bulbs» clearly exceeds the number of cases «without frost», a condition which contrasts with the remaining two months. This result shows the right tendency, although it should be remembered that the polyethylene-bulbs of the net radiometer have only a limited emission-capacity. They are therefore not so much influenced by frost-formation due to radiative cooling as by fog deposits due to the freezing of supercooled fog droplets, in any case much more so than a flat snow surface. Another source of heat is the deposition of snow at higher temperatures. Two heavy (by Antarctic standards) snowfalls in May and June (see fig. 7) occurred at air temperatures not far below freezing. A layer of 10 cm of fresh snow deposited at 20°C higher temperature could supply about 20 ly; the possible error in the estimate of sublimation heat therefore should be quite small.

The heat budget for the winter night demonstrates the importance of heat supply to the snow by turbulent mixing. Only about one quarter of the radiative energy loss causes the cooling of the uppermost 800 cm of snow. About half of the radiative energy loss is compensated by the flux of sensible heat due to turbulent mixing under inversion conditions, the remaining Quarter being latent heat supplied in the process of sublimation of hoar frost. The energies involved in the heat budget of the winter night are surprisingly small; the average net heat loss of the snow amounts to only about −6 ly per day. The low average daily heat loss in the winter of 1957 was caused by the strong warming in June, due to cyclonic activity and meridional advection of warm air. For the remaining three months the average heat loss would amount to about −9 ly per day. The winter of 1957 is hardly to be considered exceptional, since

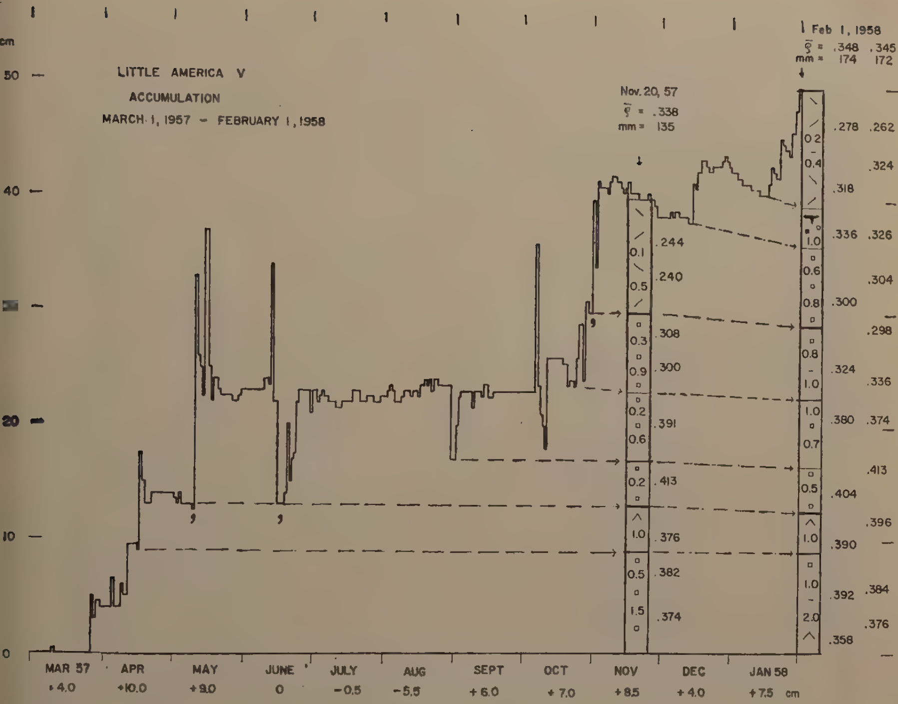

Fig. 7 — Height of snow surface (cm) above the level of March 1, 1957, at «red thread»—study site near Little America V, Antarctica. Snow stratigraphy and density distribution at this site on November 20, 1957, and on February 1, 1958.

the «Kernlose» pattern of winter temperatures seems to be typical not only of Antarctica (see H. Wexler 1959) but of the central polar regions in general. The heat preventing the development of a sharp minimum is supplied by large-scale meridional advection of warmer air in connection with cyclonic activity. This process not only increases the supply of convective heat but also reduces the outgoing net long-wave radiation by cloud layers which extend far inland.

4. The accumulation and stratigraphy of snow

Surface features of the snow cover were changing rapidly through the formation, movement and erosion of sastrugi by the wind. All research workers in polar regions were met with these difficulties and it is selfevident that only the statistical analysis of accumulation observations from a large net of stakes can lead to reliable values for the net accumulation. A dense network of snow-stakes was set up in the vicinity of Little America V and inspected at certain intervals by A.P Crary and others. The results were summarized by R.P. Goldthwait and R.L. Cameron (1960), giving the average annual accumulation at Little America as 21 cm of water. The aim of the present author's accumulation studies was the detailed picture of the building up of the snow cover and the interpretation of the stratigraphy in connection with

weather events. In doing this it was hoped to collect information, useful for the recognition of annual layers.

Near the mounting for the radiation instruments, about 270 m northeast from camp, two wooden stakes of a diameter of 5 mm were set, and their heights read daily. In addition about 6 m^2 of the snow surface nearby were covered with a wide-meshed net of thin red thread on March 1, 1957, and marked with two flags of split bamboo. This field was never disturbed, and the height of one of the flags was also measured daily. A part of the red thread was uncovered on November 20, 1957, and the stratigraphy and density of the snow studied. The pit was then closed and the surface reestablished carefully. After the return from the South Pole and shortly before leaving Little America, on February 1, 1958 an adjoining part of the red thread was uncovered again and the stratigraphy and density distribution was studied at almost the same spot. In this way it was possible to reach the level of March 1, 1957, after 265 and after 337 days, respectively, without producing any lasting disturbance.

The result is shown in fig. 7. Pronounced changes in snow level occurred in connection with storm-periods around mid May, mid June, the end of August and first half of October. The main lasting increase of the snow cover took place in fall (March-April) and in spring (October, November). The only remaining winter-snow to be found in the profile was deposited in late June on an icy crust formed by freezing drizzle on May 10 and again on June 15. The crust was exposed by the storm of June 14, which eroded away all of May's snow and that of the first half of June. The storms of August 30 and of October 7 did not erode to the same level, thus leaving about 4 cm of June's snow, which because of its fine grain and close packing has the highest density in the profile of 0.413 g.cm^{-3}. No snow from May, July, August and practically none from September remained in the profile. Only after October 10 the snow-height began to increase again. The gradual decrease of snow height through most of November and in the last days of December until mid January is mostly due to settling. On December 30 the highest temperature of -0.2°C was recorded by the ventilated thermohm and some melting took place, the signs of which were found as irregular clusters of coarse grain and ice-lenses about 12 cm below the surface of February 1, 1958. The surest means of identifying certain horizons were two thin ice-crusts: one at 21 cm below the surface of February 1, 1958, was formed during freezing drizzle on November 1, 1957, the other at 37.5 cm below the surface, formed as mentioned before on June 15 and May 10, respectively. Below the latter a few cup-crystals were found and others at the level of March 1, 1957.

The average density for the fifty cm of snow, which were deposited between March 1, 1957, and February 1, 1958, was measured as 0.348 g.cm^{-3}, the water equivalent being 17.4 cm. The pit was deepened in order to get a whole year's accumulation value. Below the red thread 8 cm of loose snow were subdivided by a thin crust, the lower 4 cm of which contained cup-crystals between 1 and 2 mm in size. The averag density of this layer was 0.321 g.cm^{-3}, and its water equivalent 2.6 cm. It consisted of late summers snow, presumably of February 1957, because it was deposited on top of another, somewhat harder snow layer of a thickness of 8 cm, 5 cm of which were characterized by signs of melting in the form of clusters of coarse grain and irregular ice-lenses and -layers. Below this level of the summer 1956/57 followed hard, fine-grained snow with a density of 0.446 g.cm^{-3}, apparently from the winter 1956. The best possible estimate for the net accumulation during the one-years period February 1957 through January 1958 is therefore 58 cm of snow with an average density of 0.345 g.cm^{-3} and a water equivalent of 20 cm.

Table 6 contains monthly values for accumulation and erosion (including settling and evaporation), summarized from daily measurements in cm of snow. The difference between both values is the net accumulation. The next two columns give the amount of snow which actually remained in the snow cover in cm of snow and in cm of water.

The latter column is compared with the results of precipitation measurements with a shielded snow-gauge (Alter type shield). Although wellknown difficulties make the results obtained with precipitation-gauges in the polar regions less trustworthy than in other climates, the comparison confirms the picture derived from net-accumulation figures. The water equivalent of the snow remaining from the single months in the profile is in excess of the measured precipitation in spring, summer and fall and shows a marked deficit in winter. The true loss due to drifting snow was presumably smaller than the 11 per cent estimated, because the most severe storms during winter were from the north to northeast, thus carrying the snow inland. The less severe winds during the rest of the year deposited more snow, than they removed. The snow carried by the prevailing winds from the southeast across the coastline into the sea was certainly only a part of the estimated loss.

The best means of identifying seasonal snow layers are combined observations of grain size, crystal structure and density. Typical of summer layers is coarse grain, low density and sometimes signs of melting as clusters of frozen-together grains and irregular ice-lenses. Typical of winter layers is fine-grained snow in close packing, leading to high density values. Thin ice-layers are more typical of winter-and spring-snow, because they indicate an immediate freezing of water on the cold surface without the possibility of trickling down. Cup crystals are rather typical of fall-layers, when the surface cools off quite rapidly and a strong water-vapour-gradient exists in the snow cover. These principles are well-known and were frequently used in the firn-regions of temperate glaciers (H. Hoinkes, 1957) and of polar glaciers. In the polar regions, where precipitation is small and the shifting of snow by the wind of great influence, the problem becomes difficult insofar, as the snow of a whole season or even a whole year might be removed from the profile in some places. Again, statistical studies with the help of certain sequences of strata preserved in the majority of the snow-profiles should lead to reasonable results (W. Vickers, 1960).

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TABLE 1

Frequency Distribution, net long-wave radiation, ly/min; hourly mean values for the winter night April 21 to August 20, 1957, Little America V, 78° 11′S, 162° 10′W

cloudiness	+ 0.04 + 0.05	+ 0.03 + 0.04	+ 0.02 + 0.03	+ 0.01 + 0.02	+ 0.00 + 0.01	− 0.00 − 0.01	− 0.01 − 0.02	− 0.02 − 0.03	− 0.03 − 0.04	− 0.04 − 0.05	− 0.05 − 0.06	− 0.06 − 0.07	− 0.07 − 0.08	− 0.08 − 0.09	n	Average net radiation ly/min
0						1	35	307	199	170	84	24	1	3	824	− 0.0354
1						3	8	45	35	10	17	5	3	1	127	− 0.0354
0 + 1						4	43	352	234	180	101	29	4	4	951	− 0.0354
2 − 9		1	11	30	45	96	178	154	112	57	22	12	4		722	− 0.0212
10 low	3	24	68	111	184	228	51	12	5						686	+ 0.0045
10 blowing snow		4	10	32	31	121	34	11	3						246	− 0.0013
10 middle + high	1	10	28	50	61	69	47	33	11	9	4				323	− 0.0018
10 all	4	38	106	193	276	418	132	56	19	9	4				1255	+ 0.0018
all cases	4	39	117	223	321	518	353	562	365	246	127	41	8	4	2928	− 0.0160

TABLE 2

Net long-wave radiation from the snow surface near Little America V, April 21 to August 20, 1957. 824 hourly mean values with cloudless skies, as a function of the recorded temperature of the instrument.

Temperature interval		*n*	Group means		
°F	°C		°F	°C	net radiation (ly/min)
≧0	≧ −17.8	5	− 1.2	− 18.4	−0.0622
− 0.1 - −10.0	−17.8 - −23.3	9			
−10.1 - −20.0	−23.4 - −28.9	24	−14.5	−25.8	−0.0568
−20.1 - −30.0	−28.9 - −34.4	84	−27.2	−32.9	−0.0452
−30.1 - −40.0	−34.5 - −40.0	224	−35.7	−37.6	−0.0394
−40.1 - −50.0	−40.1 - −45.6	290	−45.2	−42.9	−0.0310
−50.1 - −60.0	−45.6 - −51.1	179	−54.4	−48.0	−0.0285
< −60.1	< −51.2	9			
		824	−41.0	−40.6	−0.0354

TABLE 3

Daily totals of net long-wave radiation (ly/day) and daily means of cloudiness (tenths) for the winter night, April 21 to August 20, 1957. Little America V, 78° 11'S, 162° 10' W

	April		May		June		July		August	
	NR	C	NR	C	NR	C	NR	C	NR	C
1			+ 9.4	8.5	+ 14.4	6.6	+ 4.1	8.0	– 48.1	5.3
2			– 24.0	7.8	– 16.7	5.2	– 10.8	5.3	– 26.6	4.7
3			– 58.7	6.0	– 51.4	0.2	– 19.1	6.0	– 19.6	6.3
4			– 42.9	1.3	+ 0.9	5.1	– 36.0	4.0	– 15.3	9.8
5			– 17.1	4.5	– 1.8	9.2	– 40.3	0.3	– 51.7	5.5
6			– 14.0	5.7	– 43.7	4.0	– 21.3	4.3	– 11.6	6.5
7			– 17.3	7.5	– 16.9	8.8	+ 18.2	10.0	– 17.8	5.2
8			+ 30.3	10.0	– 11.0	9.5	– 37.3	2.5	– 12.4	4.1
9			– 2.9	10.0	– 28.9	6.5	– 47.4	0.2	– 28.2	2.1
10			+ 14.8	10.0	+ 4.9	5.8	– 34.0	0.1	– 2.1	6.7
11			– 10.1	10.0	+ 10.9	9.5	+ 29.0	9.5	– 28.3	4.6
12			– 46.8	6.8	– 5.0	9.9	+ 2.5	8.6	– 13.5	6.0
13			– 6.8	9.7	– 6.9	10.0	+ 8.5	9.5	– 3.8	6.0
14			– 14.7	9.1	– 22.1	10.0	– 1.2	10.0	+ 6.6	6.5
15			– 30.2	8.3	– 7.2	10.0	– 46.0	3.3	– 29.7	2.4
16			– 21.8	7.4	– 15.7	8.8	– 50.1	3.4	– 26.9	2.7
17			– 15.8	8.8	– 23.5	8.3	– 68.8	1.0	– 44.2	0.5
18			– 16.5	9.8	– 19.1	9.0	– 52.4	0.2	– 24.4	5.0
19			– 55.7	3.4	– 35.0	8.9	– 32.2	2.4	– 5.7	7.5
20			– 60.4	0.8	– 77.0	5.0	+ 4.6	6.8	– 17.8	8.0
21	+ 0.7	9.0	– 65.8	1.0	– 26.6	7.6	– 9.1	9.0		
22	– 55.3	3.4	– 63.6	0.6	– 27.6	5.3	– 25.8	8.7		
23	– 38.3	1.4	– 47.0	0.0	+ 0.4	9.5	– 23.0	9.0		
24	– 36.6	0.2	– 39.8	0.0	– 6.3	10.0	– 15.0	9.3		
25	– 37.6	0.8	– 34.4	0.0	– 22.0	5.3	– 69.0	0.0		
26	– 34.5	0.0	– 35.9	0.9	– 48.1	1.5	– 66.5	0.0		
27	– 36.3	0.2	– 45.7	0.2	– 56.9	2.8	– 40.2	3.5		
28	– 3.3	5.1	– 37.6	0.0	– 13.7	8.7	+ 4.2	8.9		
29	+ 15.3	9.1	– 34.1	0.0	– 2.4	10.0	– 22.9	7.1		
30	+ 22.1	9.5	– 23.3	0.9	– 54.5	0.9	– 27.0	5.1		
31			– 25.7	2.5			+ 3.0	9.4		
	– 203.8		– 854.1		– 608.5		– 721.3		– 421.1	
mean	– 20.4	3.9	– 27.6	4.9	– 20.3	7.1	– 23.2	5.3	– 21.0	5.3
avg. temp. °C	– 36.1		– 30.9		– 24.1		– 35.4		– 34.9	

TABLE 4

Heat balance of the snow during the winter night (sun continuously below the horizon) April 21 to August 20, 1957, Little America V (78° 11′S, 162° 10′W)

	April 21-30	May 1-31	June 1-30	July 1-31	August 1-20	April 21-August 20	%
$\Delta \bar{T}$ snow, 0-8 m, °C	− 0.9	− 1.7	+ 0.8	− 1.8	− 0.5	− 4.1	
Δ heat content snow, ly	− 149	− 281	+ 132	− 298	− 83	− 679	− 24%
Net long-wave radiation, ly	− 204	− 854	− 609	− 721	− 421	− 2809	− 100%
Convective heat, ly	+ 55	+ 573	+ 741	+ 423	+ 338	+ 2130	
Sensible heat, ly (Liljequist's method)	+ 120	+ 334	+ 282	+ 418	+ 286	+ 1440	+ 51%
Sublimation heat, ly	− 65	+ 239	+ 459	+ 5	+ 52	+ 690	+ 25%
Sublimation grams per cm²	− 0.10	+ 0.35	+ 0.68	+ 0.01	+ 0.08	+ 1.02	

TABLE 5

Formation of frost on the net radiometer (number of cases) during the winter night 1957, at Little America V

	April 21-30	May 1-31	June 1-30	July 1-31	August 1-20	April 21-August 20
no frost	17	37	30	52	32	168
slight frost	15	41	25	32	27	140
heavy frost	6	19	20	7	13	65
clear ice	0	4	3	0	0	7
frost / no frost	1.24	*1.73*	*1.60*	0.75	1.25	380

TABLE 6

Accumulation, erosion and precipitation at Little America V, February 1957 to January 1958.

	Accum.	Settling evaporation erosion	Net Acc.	in the snow profile on Feb. 1, 1958		precipitation measured	with respect to precipitation gain	loss
	cm of snow				cm of water			
Feb. 1957	—	—	—	8.0	2.60	1.70	0.90	
March	+ 7.0	— 3.0	+ 4.0	4.0	1.45	1.67		0.22
April	18.5	8.5	+ 10.0	8.5	3.32	0.91	2.41	
May	39.0	30.0	+ 9.0	—	—	5.86		5.86
June	28.5	28.5	0.0	4.0	1.61	3.81		2.20
July	3.5	4.0	— 0.5	—	—	0.94		0.94
August	5.5	11.0	— 5,5	—	—	0.48		0.48
September	9.0	3.0	+ 6.0	1.0	0.38	0.81		0.43
October	34.0	27.0	+ 7.0	11.5	4.25	3.25	1.00	
November	20.5	12.0	+ 8.5	7.0	2.10	1.47	0.63	
December	8.5	4.5	+ 4.0	3.5	1.19	0.56	0.63	
January 1958	12.0	4.5	+ 7.5	10.5	3.10	1.09	2.01	
Total 11 month	+ 186.0	— 136.0	+ 50.0	50.0	17.40	20.85	+ 6.68	— 10.13
Total 12 month				58.0	20.00	22.55	+7.58	— 10.13 — 2.55

CHEMISTRY OF ICE, SNOW AND OTHER WATER SUBSTANCES IN ANTARCTICA

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1. Introduction

To determine the composition of various types of water substance in Antarctia is not only important from the viewpoint that it offers materials for drawing a whole scientific picture of this region of the globe but is also of special significance at the point that the obtained information is helpful to complete our knowledge of the transportation of various chemical species on the whole earth and the ways and mechanisms through which such a transportation is actualized. This initiated us to undertake the present research.

All the samples were collected by the members of the Japanese Antarctic Research Expedition Team at locations not far away from Showa Base, the headquarters of the Japanese Antarctic Research Expedition, 68°30′S, 40°30′E. The following kinds of elements were determined.
Na, K, Mg, Ca, Sr, Cl, SO_4, I, P and As (*).

2. Results and characteristics of waters

2.1. *Pool water of East Ongul Island*

There are found many fresh-water pools on East Ongul Island some of which remain unfrozen under icecover during the winter time. Three samples of these waters were collected by T. Torii, a member of the Japanese Antarctic Research Expedition Team, in February, 1957, to be transported to the author's laboratory. The results of chemical analyses listed in Table 1. are characterized by two points: 1) Different from ordinary natural waters the ratio Na/Cl in them is lower than that in sea water. 2) The ratio SO_4/Cl in them is also lower than that in sea water.

On these two characteristics discussions will be made later.

(*) Na, K, Ca and Sr. determined by N. Kawasaki by using the flame photometric method proposed by K. Sugawara, T. Koyama and N. Kawasaki, *Bull. Chem. Soc. Japan*, **29**, 683 (1956).
Mg determined by N. Kawasaki by the usual EDTA titration method.
Cl. Samples rich in Cl were determined by Mohr's method, while samples poor in Cl by K. Terada in using S. Utsumi's thiocyanate method, *J. Chem. Soc. Japan*, **73**, 835 (1952).
SO_4 Samples rich in SO_4 were determined by $BaSO_4$ gravimetric method, while samples poor in SO_4 by E. Kamata, using the barium chloranilate method by R.J. Bertolacini and J.E. Barney, *Anal. Chem.*, **29**, 281 (1957).
I. determined by K. Terada by using the spectrophotometric method proposed by K. Sugawara, T. Koyama and K. Terada, *Bull. Chem. Soc. Japan*, **28**, 494 (1955).
P and As. determined by S. Kanamori by spectrophotometric methods invented by K. Sugawara and S. Kanamori (unpublished).

2.2. *Snow*

Two samples of snow have been examined. As listed in Table 2., the ratio Na/Cl in sample no. 1 is close to but slightly higher than that in sea water, while the ratio in sample no. 2 is evidently higher than that in sea water as usually found in samples of rain and snow which have been collected in other regions of the globe. At this point snow differs clearly from the pool waters.

On the other hand,the ratio SO_4/Cl in snow as in the pool waters is lower than that in sea water.

Another characteristic to Antarctic snow is its high content of phosphorus.

2.3. *Packice*

Two examples of the results of analyses of pack ice samples are listed in Table 2.

Pack ice is near pool waters in nature and characterized by lower values of both ratio Na/Cl and ratio SO_4/Cl.

3. Discussions on the characteristics of Antarctic water subsatnces

Ratio Na/Cl

Among various kinds of natural waters, the sea water is lowest in the ratio Na/Cl, its value being 0.856 as expressed in equivalent. Thus, 1.87 is the average of the ratio Na/Cl in 220 representative rivers and lakes in Japan and 1.6 is the average for world rivers, while the average for thermal springs in Japan is 0.95 (see Table 3). The ratio in rain and snow stand intermediately between the sea and terrestrial waters. Besides, the ratio in rain gradually changes from sea coast towards inland and from a low level upwards. Thus, the ratio in samples which were collected at a coastal area is found close to the sea value, and as the point of collection moves away from the coast towards inland and away from sea level to higher levels, the ratio is found to gradually increase. This suggests that the primary source of the atmospheric salt and consequently the salt in rain and snow is sea spray which is constantly emitted into the air through wave breaking. Then the question is: How can we explain the gradual deviation of the composition of the salt in rain mentioned above? A topic of argument. The idea is divided among various experts. The present author accepts that there exist many other sources of salt which are active in modifying to a greater or smaller degree the original composition of the emitted sea salt. At the same time he is convinced that the most important factor causing the deviation of salt composition is the post-wave-breaking fractionation so he termed 1a, b, c). The fundamental of this idea is that the emitted sea spray is subjected to evaporation, while suspended in the air, followed by a successive fractionation and separation of different kinds of salt particles. They differ in composition from one to another and differ in stability in surviving in the air. Some of the particles fall down earlier from the air, others are easily taken in rain and snow and some others are likely to be eliminated by some objects on the path because of their adhesiveness. Thus through the elimination of less stable components the gross composition of the atmospheric salt gradually deviates from the original composition.

The stability of elements is assumed to increase in the following sequence

$$Cl < Na < Mg,\ K < SO_4,\ Ca < Sr < I$$

As seen from this sequence, Cl is the least stable component and it is easily understood

that the ratio Na/Cl in rain and snow is greater than the coresponding value in sea water and that the ratio tends to increase horizontally and vertically with distance away from the sea.

Returning to our main subject of the low ratio Na/Cl of the pool waters of East Ongul Island, we must find an explanation to it.

It is evident that the primary source of the water of the pools is the snow which fell on the surrounding region. At the same time it is quite probable that an additional source of salt is the sea spray which was directly transported from the sea in the vicinity.

Now, let us imagine that the author's theory of post-wave-breaking fractionation is valid and that the first fraction of elimination enriched by Cl constitutes the transported spray.

Then it appears rather natural to find the composition of the pool water is sufficiently affected by this transported spray so that its Na/Cl becomes low as actually observed. The situation, however, is not so simple. Some other components which stand higher in the stability sequence are found already enriched in the pool water. In fact, as seen in Table 1, the enrichment coefficient, $(M/Cl)_{pool\ water}/(M/Cl)_{sea\ water}$ is found to be greater than unity for components such as Mg, Ca, Sr and I.

Then it is only after this complicacy is solved that the validity of the proposed explanation can be positively accepted.

Aside from this explanation, the process of synfractionation of sea salt is another important factor causing the observed low ratio Na/Cl[8] .

Synfractionation of sea salt is an experimentaly established fact which the author reported in detail at the International Oceanographic Congress in New York, last September. The fundamental is that when droplets are formed through collapse of sea foam to be emitted into the air, the composition of salt in the droplet differs already from that of the sea water from which the spray came. This shows that different salt components escape with different eases at the burst of bubles to become the components of the emitted spray. The relative ease can be expressed in terms of enrichment coefficient, $(M/Cl)_{spray}$ $/(M/Cl)_{sea\ water}$, where $(M/Cl)_{spray}$ and $(M/Cl)_{sea\ water}$ stand for the ratios of one component to Cl in spray and sea water respectively. The enrichemnt coefficient was experimentally proved to increase in the following sequence

$$Na < Cl,\ Ca,\ Sr < SO_4\ Mg < I.$$

Most remarkable is that the ease of escape is greater for Cl than for Na with resultant spray more enriched by Cl than Na.

Then the transportation of spray from such synfractional process is quite probable to considerably contribute to the observed low ratio Na/Cl in the pool water.

The low ratio Na/Cl in packice samples is also understandable when we consider that while the ice was floating and growing on the sea, it must have been exposed to the effect of the sea spray which was produced by synfractionation process.

Ratio SO_4/Cl

The SO_4 problem regarding the Antarctic water substances in which the ratio SO_4/Cl is nearly always lower than the corresponding value in sea water is more difficult to solve.

All possible factors which modify this ratio in waters work to increase it. Thus starting with the ratio in sea water, 0.1007, the ratio ranges 0.115 —3.2 in meteoritic waters and the average in 220 representative rivers and lakes in Japan is 1.35 (see Table 4). One example of exception is Dead Sea with a value 0.0019 in which the preferential precipitation of $CaSO_4$ by evaporation is the cause of the low value. Even synfractionation process acts to increase the ratio. The only one conceivable idea is of the first elimination fraction of the post-wave-breaking fractionation, in which

the ratio SO_4 Cl is likely to be lower than in sea water. However, we must face a similar difficulty again which we met in the previous section in the discussion on the Na-Cl relation. How can we explain that except SO_4 the whole picture of the composition of salt is rather to be taken for indicating a later stage of fractionation? Especially important is that the SO_4 shortage is to every kinds of land and meteoritic waters in Antarctica so far tested. The final solution of this problem can be given only through an exhaustive investigation.

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TABLE 1

Chemical composition of pool waters in the Ongul Island

		No. 1	No. 2	No. 3	average
Cl	mg/l	135.7	136.$_3$	204.$_2$	158.$_4$
Na	mg/l	74.$_0$	74.$_0$	85.$_0$	77.$_7$
K	mg/l	2.4$_6$	3.3	5.5	3.75
Mg	mg/l	9.6$_5$	13.4	13.$_7$	12.$_3$
Ca	mg/l	7.7	14.0	11.0	10.$_9$
Sr	mg/l	0.13	0.36	0.21	0.23
SO_4	mg/l	36.2	16.4	25.2	25.9
I^-	µg/l	1.9	1.9	0.5	1.4
IO_3	µg/l	2.8	2.9	5.0	3.6
I_t	µg/l	4.7	4.8	5.5	5.0

Na/Cl	in equiv.	0.80	0.84	0.64
K/Cl	in equiv.	0.017	0.022	0.024
Mg/Cl	in equiv.	0.21	0.29	0.19$_6$
Ca/Cl	in equiv.	0.10	0.18	0.09$_5$
Sr/Cl	in equiv.	0.0007$_8$	0.0018	0.0008$_3$
SO_4Cl	in equiv.	0.19$_8$	0.089	0.091
It/Cl	in equiv.	0.0000097	0.0000098	0.000007$_5$

Enrichment Coefficient			
Na	(0.9$_3$)	(0.9$_7$)	(0.7$_4$)
K	(0.9$_3$)	1.2	1.3
Mg	1.1	1.5	1.0
Ca	2.7	4.9	2.6
Sr	2.3	5.2	2.5
SO_4	2.0	(0.8$_6$)	(0.8$_8$)
It	13	13	10

TABLE 2

Chemical composition of snow and pack-ice in Antarctica

	snow		pack-ice	
	1	2	1	2
Cl mg/l	$413._7$	8.3	$995._6$	3.4
Na mg/l	240	5.3	525	1.5
Mg mg/l	$20._4$	0.61	$65._8$	0.15
Ca mg/l	8.9	0.42	$21._2$	0.19
SO_4 mg/l	46	0.8	119	1.2
I μg/l	$10._5$	4.1	8.0	1.4
P μg/l	81	480	250	1500
As μg/l	4.1	2.6	3.2	4.8
Enrichment Coefficient				
Na	1.0_2	1.1_7	0.95	0.78
Mg	0.72	1.1	0.96	0.67
Ca	1.00	2.4	1.0_1	2.7
SO_4	0.80	0.7	0.85	2.6
I	9.5	190	3.0	160
P	53	16,000	63	120,000
As	38	1,20	12	5,40

TABLE 3

Na/Cl in various kinds of natural waters

Sea		0.86
River and lake	(Japan)	1.87
River	(World)	1.60
Rain and snow	(Japan)	1.30
Pool	(Antarctica)	0.76
Snow	(Antarctica)	0.94
Pack-ice	(Antarctica)	0.74
Thermal springs	(Japan)	0.95
Dead Sea	(Jordan)	0.26

TABLE 4

SO_4/Cl in various kinds of natural waters

			Cl mg/l	SO_4 mg/l	SO_4/Cl in equivalent
Mist	Hallde, Norway		3.56	0.57	0.115
Fog	Coast of Nova Scotia and Massachusetts		—	—	0.37-1.23
Rain	Japan				0.37
Snow	Japan				3.2
River	Japan		5.8	10.6	1.35
Sea water			19,000	2,650	0.1007
Pool water		1.	134.7	36.2	0.19_8
	Antarctica	2.	136.3	16.4	0.089
		3.	204.2	11.0	0.091
Newly fallen snow		1.	$413._7$	4.6	0.082
	Antarctica	2.	8.3	0.8	0.07
Pack-ice	Antarctica	1.	$995._6$	119	0.088
		2.	3.4	1.2	0.26

DEPLACEMENT DU GLACIER DE L'ASTROLABE

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RÉSUMÉ

Après un exposé des mesures effectuées et de la méthode de dépouillement utilisée, nous donnons les élements du déplacement des balises implantées sur le Glacier de l'Astrolabe en 1958, trajectoire, vitesse et direction.

Après discussion de la précision de ces résultats, l'accélération longitudinale du glacier et le profil transversal des vitesses superficielles sont exposés.

Aucune variation saisonnière marquée n'est mise en évidence au cours de l'année à part quelques à-coups davantage liés au régime des vents qu'à celui de la glace de mer.

Enfin, nous montrons le faible débit de glace de la calotte antarctique et sa régression dans la partie côtière de Terre Adélie.

INLAND ICE MOVEMENT IN MAC ROBERTSON LAND, ANTARCTICA

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Summary

Details are given of ice thickness and surface velocity measurements along an east-west line 12 km. inland from the coast of MacRobertson Land. Approximately 68×10^5 kg. of water equivalent of ice flows across each metre of this line annually, and 38×10^5 kg. of water equivalent of ice reaches each metre of this part of the coast-line each year.

1. Introduction

In 1957, the glaciologist with the Australian National Antarctic Research Expedition, M. Mellor, established a line of ice flow stakes a few km north of the Framnes Mountains, about 12 km inland from Mawson.

The displacement of these stakes was observed at subsequent dates, to obtain a measure of the flow rate of the ice at the surface. In the autumn of 1958, ice thicknesses along the same line were measured by seismic and gravimetric methods, in an attempt to correlate the thickness and bedrock configuration with the rate of flow of the ice.

The data obtained in 1958 enable a quantitative estimate to be made of the rate of sheet flow near the margin of the Antarctic ice cap in the Mawson region. The rate of movement is obviously influenced by the nearby ranges, but nevertheless the results probably indicate the quantity of ice movement in similar near-coastal regions of the Antarctic.

A brief outline of the measurements has been given by Mellor (1959).Crohn (1959) has also made an estimate of ice movement, based on rates of flow close to Mt. Henderson, a large rock exposure. He obtained a figure a little less than half that found by these later calculations.

2. General Setting

The ice sheet in this area rises steeply from sea level, with numerous valleys and crevassed domes, so that at the northern end of the Framnes Mountains it has an altitude of about 500 metres. The transition from the zone of ablation to the zone of accumulation takes place at altitudes between 700 and 900 metres, some 20 km inland, so that the line of flow stakes is entirely within the ablation zone. All the stakes were set in hard bubbly ice except for nos. 0, 3 and 6, which were on lee slopes,and set in névé containing numerous ice layers and blebs.

3. Measurements

The base line for the flow measurements is an east-west line between the Fischer Nunatak (just south of Mt. Henderson) and the highest point of the Casey Range.

Twenty-three stakes, situated at irregular intervals along the eastern 26 km of this line, were used for the measurements.

Movement of each stake was measured in January and autumn of 1958 by determining the original base line by triangulation using the two reference points, and measuring the offset to each stake. During the autumn measurements, the distance of each stake along the line from the summit of the Fischer Nunatak was also measured, as a preliminary to measuring the east-west component of flow.

The offset distances were measured to the nearest 6 inches (15 cm) (although the actual accuracy would be rather less than this, approximately $\pm$ 50 cm) and the longitudinal distances to the nearest yard (1 metre). The time interval between the establishment of the line and the autumn measurements ranged from 305 days at the eastern end to 350 days at the western end.

Ice thicknesses were measured by seismic reflection and gravimetric methods (Jesson, 1959). The values obtained by the seismic method were used as control on the gravity work. 8 seismic and 22 gravity stations were established along the flow stake line, and two short gravity cross-traverses were also made.

The rate of ice movement is shown in Table 1, and the ice thickness and surface altitude along the flow stake line in Table 2; Tables 3 and 4 give ice thicknesses and altitudes for the two cross-traverses.

It appears that the rate of movement depends more on the position in relation to the nearby ranges than on the bedrock topography. Well defined ice streams flow between Mt. Henderson and the Masson Range, and between the David and Casey Ranges; the latter stream develops into the Forbes Glacier. Although there is quite a marked bedrock channel between Mt. Henderson and the Masson Range, the northward flow rate is not very great; the surface topography indicates a barrier to the north of the flow stake line here, and there is probably a considerable westward component to the flow between poles 2 and 6.

4. Volume of ice flow

From the figures of Tables 1 and 2, an estimate can be made of the total volume of ice moving annually across the flow stake line.

The actual state of movement of ice is complex, and has not yet been given an exact mathematical treatment, although several workers, e.g. Nye (1957) and Vialov (1958) have developed expressions for various types of flow. For this reason, no more than a very approximate estimate is justified in the present case, because so little is known of ice thicknesses and bedrock topography on either side of the line (with consequent uncertainty as to the vertical velocity profiles) and because in places there are indications of a component of flow parallel to the line.

If the ice along the length of the line is considered to be divided into a series of slabs, each corresponding in width to the distances between adjacent flow poles, then the vertical cross-sectional area (parallel to the flow stake line) of these slabs can be found by interpolating ice thicknesses, and the volume flowing across the line by multiplying this area by the mean of the speeds at the appropriate poles. This procedure will, of course, give the maximum possible value for the volume.

By this means, it is found that 5750 cubic metres of ice flows daily across a line 27160 metres long. If the average specific gravity is taken as 0.876, this volume is equivalent to 1839×10^{6} kg of water a year, i.e. an average of 68×10^{5} kg of water moves annually across each metre of the flow stake line.

By taking the average annual ablation between the line and the coast, which is about 12 km to the north, as 25 cm of water (an arbitrary estimate, derived from ablation measurements at several altitudes, and a surface profile running inland from

Mawson) the annual water loss by ablation from the surface of a 1 metre strip between the line and the coast is 30 × 10^5 kg i.e. an average of 38 × 10^5 kg of water in the form of ice moves across each metre of this part of the coastline each year.

TABLE I

Surface flow rates.

Stake	Distance from Fischer Nunatak	Movement per year	Movement per day
0	1273 metres	10.4 metres	2.8 cm
1	2456	13.7	3.7
2	3632	14.8	4.1
3	4752	17.2	4.7
4	5859	19.2	5.2
5	6285	22.2	6.1
6	8525	15.2	4.2
7	9472	10.0	2.7
8	10523	11.3	3.1
9	11684	13.5	3.7
10	12599	15.7	4.3
11	13752	15.2	4.2
12	14359	15.6	4.3
13	15517	19.7	5.4
14	18275	21.3	5.8
15	20454	15.1	6.9
16	21708	25.7	7.1
17	22934	33.0	9.0
18	24353	36.7	10.1
19	25594	33.5	9.2
20	26626	35.3	9.7
21	26933	38.8	10.6
22	27161	43.3	11.9

TABLE II

Ice thicknesses and surface altitudes, main east — west traverse

Point	Distance from Fischer Nunatak	Surface altitude	Ice thickness
SP27	1273 metres	529 metres	422 metres
SP28	2882	516	550
SP29	4492	501	531
SP30	6058	508	623
G7	6966	485	587
G6	7771	466	596
G5	8575	456	589
G4	9380	449	569
G3	10185	453	545
G2	10989	451	494
G1	11794	458	417
SP31	12599	447	404
G8	13403	434	390
G9	14208	443	402
G10	15013	440	402
G11	15817	456	385
G12	16622	449	387
G13	17427	439	458
G14	18231	458	495
SP32	19036	454	459
G15	19841	431	436
G16	20645	429	438
G17	21450	477	355
G18	22255	487	333
G19	23059	497	317
G20	23864	487	332
G21	24669	492	292
SP33	25473	497	268
G22	26601	532	271

TABLE III

Ice thicknesses and surface altitudes, cross-traverse B. Stations at ½ mile (805 metre) intervals

Point	Surface altitude	Ice thickness
(South) G25	574 metres	275 metres
G24	538	264
G23	519	264
SP33	497	268
G26	465	337
G27	468	252
G28	439	242
(North) G29	394	264

TABLE IV

Ice thicknesses and surface altitudes, cross- traverse C. Stations at ½ mile (805 metre) intervals

Point	Surface altitude	Ice thickness
(South) G37	620 metres	490 metres
G36	590	483
G35	554	489
G34	545	508
G33	528	478
G32	509	444
G31	485	429
G30	470	421
SP31	447	404
G38	420	387
G39	380	383
G40	353	395
G41	336	435
(North) G42	320	473

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ZONES OF SNOW ACCUMULATION IN EASTERN ANTARCTICA

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SUMMARY

Researches of the Soviet Antarctic Expedition have established the character of distribution and dynamics of the snow cover in Eastern Antarctica. An attempt is made below to divide this part of the Antarctic Continent into separate areas by different processes of snow accumulation.

In the eastern part of the Antarctic Continent the following large zones of snow accumulation can be distinguished, the character and intensity of atmospheric precipitation and dynamics of snow cover in which differ quite substantially :

1 — *Central zone*, including the inner territories of the Antarctic Continent, with small values of atmospheric precipitation, which fall only as snow and rime at very low temperatures. Here a loose snow cover predominates, owing to low wind velocities. Only fragmentary stretches and patches of hard sastrugi are found, their origin being associated with cyclonic invasions or with insignificant roughnesses of relief, which create local run-off winds.

2 — *Peripheral zone of catabatic winds*, occupying the slope of the Antarctic ice sheet in a belt up to 700-800 km wide. In this zone the snow cover is formed under the influence of cyclonic and extremely stable catabatic winds mainly at the expense of cyclonic sediments. Characteristic is an extremely uneven distribution of the snow cover, its great density and hardness, a great number of sastrugi and other forms of accumulation-deflation microrelief.

3 — *Antarctic Coastal zone*, embraced the outer, steeper part of the slope of the ice sheet, measuring from several tens to over 100 kilometres in width. In this zone a regular increase of deposited sediments towards the coast is recorded, up to the belt of maximum accumulation, inclusively.

4 — *External oceanic zone*, located beyond the sphere of influence of run-off winds. It includes Antarctic islands and the outer parts of shelf glaciers; it is characterized by a domination of a cyclonic regime and abundant precipitation.

A special place in respect to snow accumulation belongs to «oases», which very often have no continuous snow cover throughout the entire year.

RÉSUMÉ

Par les investigations de l'expédition Soviétique Antarctique furent révélées les lois principales de la distribution et de la dynamique de la couverture de neige dans l'Antarctide Orientale. Ceci permet de procéder à l'expériment, portant sur une division en régions des processus d'accumulation de neige et de la dynamique de la couverture de neige sur la surface du champ de glace antarctique.

Nous distinguons dans l'Antarctide Orientale, les zones suivantes d'accumulation de neige, dans lesquelles le caractère et l'intensité de l'accumulation des sédiments solides et la dynamique de la couverture de neige diffèrent essentiellement :

1. *La zone centrale*, qui embrasse les territoires internes de l'Antarctide, à petites valeurs absolues de précipitations atmosphériques en forme de neige et de givre, avec prédomination d'aires à couverture de neige meuble, dû aux petites vitesses du vent, à bandes fragmentaires et à parcelles de « zastrougs » dures, liés aux intrusions des cyclones ou bien aux accidents du relief, causés par des vents locaux de déversements. 2. *La zone périphérique des vents de déversements*, qui occupe la pente de la coupole de glace antarctique par une bande à largeur jusqu'à 700-800 km. La couverture de neige se forme sous l'action des vents de déversement et cycloniques au compte des précipitations de type principalement cyclonique. Caractéristique est une distr-i nution extrêmement irrégulière de la couverture de neige, sa grande densité et dureté, la présence d'une grande quantité de « zastrougs » et d'autres formes de microrelief accumulatif-deflationné. Dans la limite de cette zone, l'on distingue les ceintures de transition, de déversement, d'intense accumulation, d'accumulation de fissures, et d'accumulation près des barrières. Les valeurs absolues des sédiments déposés sont considérables. 3. *La zone océanique extérieure*, située au dela de l'influence des vents de déversement et caractérisé par une prédomination du régime cyclonique et par des précipitations abondantes. Une place à part est occupée par les « oasis » antarctiques.

The natural conditions obtaining in Antarctica today are determined by: 1) the continent's single-Pole geographical position in the heart of a gigantic ocean ring ; 2) the ice sheet with special physical properties of its own, which covers almost the entire continent; and 3) the great absolute elevation and size of the continent.

The flat-bulging ice sheet, rising in the central areas to 3,500-4,000 metres above sea level and extending for 4,000 kilometres in diametre, is the major megaphromic relief of Eastern Antarctica. The profile of the ice sheet's surface is nearly that of a semi-elipse with a very flat central part and a considerably steeper slopes along the periphery. This shape of the surface is due chiefly to the properties of the moving ice itself. Beneath the ice cover there is very broken mountainous land, but because of the great thickness of the ice (1,000-4,000 metres), the surface reflects, and very faintly at that, only the major prominences of the relief beneath the ice. For example, the huge mountain ranges beneath the ice in the region of Sovetskaya Base and the Pole of Inaccessibility are mirrored on the surface by a vast but very flat tableland that has been named Plateau Sovetskoye. A deep depression, stretching for more than 1,000 kilometres from Olaf Pruds Gulf to the Pole of Inaccessibility is traced on the surface in the shape of a broad hollow with gentle slopes. The unevenness of the relief beneath the ice thus complicates the morphology of the surface of the ice sheet but does not determine it.

A law-governed change of geographical conditions, linked up with changes in height above the sea level and in distance away from it are observed as one rises up the slope of the ice sheet from the coast deep inland. Data obtained by the Soviet Antarctic Expedition in 1956-1959 show that the territory of Eastern Antarctica may be divided into concentrically arranged geographical zones that substantially differ from each other with regard to many natural phenomena and processes. We single out four such zones: central zone, zone of catabatic winds, zone of the Antarctic coast, and external ocean zone.

1. The *central zone* embraces the highest but least broken central part of the ice sheet of Eastern Antarctic, with the lower border situated at an elevation of 2,800-3,000 metres above sea level. Vostok, Komsomolskaya and Sovetskaya Bases as well as the Pole of Inaccessibility are situated within this zone. The relief is extremely monotonous with very small slope angles (10^{-3}). The features of the climate are very low negative temperatures throughout the year (in summer from −22°C to −40°C, in winter from −40°C to −88°C), the predominance of clear, cloudless weather, small wind velocities, very dry air and little precipitation. Precipitation forms either directly on the surface of the snow cover in the shape of hoar-frost (up to 50% of the total) or in the shape of tiny pole-like crystals of ice from the «snowy haze» forming in the layer of intensive subice inversion. Clouds producing snowfalls appear during cyclones blowing from the oceanic regions. The annual precipitation at Komsomolskaya Base amounts to 20-25 mm, and at Vostok and Sovetskaya bases—about 25-35 mm (Bugayev, 1960). According to other investigators, this value is much bigger. On the basis of the snow thickness in a pit, V. M. Kotlyakov (1959) determined the mean annual accumulation of precipitation at Komsomolskaya Base as 80 mm for the past 30 years, and at Vostok-1 Base as 102.5 mm a year for a period of 24 years. N. P. Rusin (1959) has calculated that through condensation the inland regions of Antarctica receive about 30-40 mm of precipitation a year. It will therefore be no mistake to surmise that the mean annual accumulation of precipitation throughout the whole of the central zone is close to 50 mm.

The mean monthly wind velocity in the central zone rarely exceeds 4-5 metres per second, and as a consequence of this the snowdrift is relatively small and the ice cover is more even and spreads in a looser layer than along the periphery of the continent. According to observations made at Komsomolskaya, the hardness of the

Fig. 1 — Scheme of geographical zones in Eastern Antarctics 1- Scientifical stations: Mirny (USSR), Pionerskaya (USSR), Vostok-1 (USSR), Komsomolskaya (USSR), Vostok (USSR), Sovietskaya (USSR), Polus Nedostupnosti (USSR), Dobrovolsky (Poland), Lazarev USSR), Sheakleton (Great-Britain), Maudheim (Norwegia), King Boduen (Belgium), Sova (Japan), Mawson, (Australia), Davis (Australia), Wilkes (Australia), Luis Isl. (Australia), Dumont-Durville (France), Sharko (France), Amundsen-Scott (USA). 2- Central zone of Antarctic glacier cover; 3- Zone of catabatic winds; 4- zone of Antarctic littoral; 5- Outer oceanic zone.

snow cover fluctuates from 0.30 to 0.90 kg/cm², while in exceptional cases it rises up to 4.5 kg/cm² with a density of 0.36-0.40 gr/cm³ (Kotlyakov, 1959). The conversion of snow into ice follows the recrystallization pattern and takes place at a depth of 120-150 metres. Despite the small precipitation, snow accumulates systematically in the central zone (there is no melting or evaporation, while transportation of snow by the wind acquires some importance solely in a narrow strip along the periphery) and to ensure a credit and debit balance of the ice sheet there must be a run-off of ice from the centre to the periphery. Aeolian forms of microrelief of the snow cover occur almost throughout the whole of the central zone, but their scale is small, the snow ridges are much less harder than in the neighbouring catabatic wind zone, and the snow cover is much less dynamic. Relatively big and hard ridges occur fragmentarily and are usually linked up either with sections where the surface slopes more steeply and give rise to very high-velocity local catabatic winds, or with the penetration of separate powerful cyclones inland from the direction of the ocean.

2. The *zone of catabatic winds* occupies the slopes of the ice sheet of Eastern Antarctica, girdling its central part with a strip 700-800 kilometres wide, its elevation ranging from 1,000-1,200 to 2,800-3,000 metres above sea level. Strong catabatic winds that blow all year round make up the most characteristic feature of this zone. Rising far inland, these winds gradually pick up speed as they come down to the shore along the increasingly steep slope and achieve storm force on the coast. The uneven relief in the path of the catabatic winds causes local deviations in direction and velocity but does not change the general picture of the run-off of cooled masses of air from the central regions of the continent to the coast. The influence of cyclones, advancing from the ocean and bringing clouds and snowfalls and accompanied by furious blizzards, is already considerably greater in this zone. Here precipitation amounts to approximately 150-250 mm a year. The strong cyclonic and catabatic winds, replacing each other almost uninterruptedly, lead to extreme unevenness in the distribution of the snow cover, create a sharply intersected accumulation-deflation microrelief on the surface of the snow, are the cause of the great hardness and density of the snow cover, call forth violent ground winds, blizzards and what is known as «snow haze», and transport a large quantity of snow outside the zone. Pionerskaya Base is situated in this zone, closer to its upper boundary. Measurements taken by the author showed that the annual accumulation of snow at the base in 1956 was about 160 mm. In the next two years there was greater precipitation and somewhat more snow accumulated (200-250 mm). The mean annual wind velocity at Pionerskaya Base reaches 10-11 metres per second, and in 98% of the cases the direction of these winds is in the range of east to south-south-east. Violent winds cause considerable movement of snow. For example, the mean intensity of the snow drift for July 1956 was 300-400 gr/cm² per hour, with a mean wind velocity of 11.6 metres per second for the month. During blizzards, when wind velocity was 15-18 metres per second, the intensity of the snow drift on the surface was 1,200-1,600 gr/cm² an hour, dropping to 1/10 at a height of 0.5 metres and to 1/50 of this quantity at a height of 1 metres above the surface. The reaccumulation of precipitated snow is extremely intensive. Detailed observations along the profile of the surface (measurements were taken at 150 points daily, with the exception of days when no noticeable change in the condition of the snow surface was observed, for a period of 6 winter months—from June to November, 1956) showed that the mean height of the accumulated snow was 203 cm, of which 186 cm were again dispersed and only 27 cm remained and merged with the stable snow sheet. If we take the quantity of accumulated snow as 100%, we shall find that 91% disperses and 9% remains. At some of the stakes the quantity of snow that accumulated in the course of these six months amounted to 350-450 cm, of which 300-350 cm were dispersed. The daily fluctuation of the height of the snow

sheet at some of the stakes was 50-70 cm. The snow sheet is especially dynamic during the winter months. For instance, in August and September 1956, the mean accumulation at 150 points along the profile amounted to 134 cm, 130 cm were dispersed and the snow sheet increased in height by only 4 cm or 2% of the initial accumulation. The snow sheet distributes very unevenly on the surface. Mosaic alternation of sections each with a snow sheet of a different age, with different physical properties and different forms of accumulative and deflative microrelief is a characteristic feature of this zone. We can single out snow covers of at least three types that occur simultaneously on the surface: old compact snow, armoured with radiation-wind crusts (hardness from 4-6 to 10-12 kg/cm², in some cases—up to 30-40 kg/cm²; density 0.45-0.52 gr/cm³) a young snow sheet in the process of formation (hardness from 2 to 6 kg/cm², density 0.42-0.45 gr/cm³) and ephemeral accumulations of loose snow in the shape of moving dunes, banks and drifts (hardness 0.2-2.5 kg/cm², density 0.37-0.42 gr/cm³). Observations made by the author on October 18, 1956 in the region of Pionerskaya Base showed that in a radius of two kilometres about 24% of the area was occupied by old snow armoured with radiation-wind crusts, 43 % by a newly formed snow cover and 33% by ephemeral accumulations of loose snow in the shape of drifts and dunes complicated by young ridges.

Among the forms of the accumulative-deflative snow micro-relief we can distinguish some that owe their rise to cyclonic blizzards and also others that had taken shape during ground winds and low storms brought on by catabatic winds. The cyclonic series of forms include vast strips of drifts extending in chains perpendicularly to the dominant direction of the wind. These strips are from two to six kilometres wide and alternate with sectors that are devoid of accumulations of young blizzard-brought snow. From the air these strips of snowdrifts resemble broad and flat snow waves. These «waves» are not immobile, and move slowly in the direction of the wind. We observed and measured a system of such «waves» along a 150-kilometre profile in the direction from Mirny to Pionerskaya. They occur not only where the conditions of the relief facilitate but also on perfectly even and straight slopes. The formation of this type of «waves» during blizzards is evidently linked up with the considerable turbulence of the snow-wind current in cyclones and with its undulating nature. Large elongated banks 50-100 metres long, 10-20 metres wide and up to 1-2 metres high, with 100-150-metre wind-blown hollows between them, form during violent cyclone-type blizzards. These banks take shape chiefly during autumn blizzards that are accompanied by storm winds. As a result of this we observe a very dense primary packing of snow, while the sharp drop in temperature that usually follows when cyclonic winds give way to catabatic winds causes the snow to harden swiftly. During the winter months predominance is taken over by catabatic winds of a lesser violence and these are no longer able to destroy the banks and only model them into the shape of ridges. The banks that survive on the surface until the onset of daylight are covered with a radiation-wind crust, which to all practical purposes safeguards them against further dispersal. The banks may be preserved in this shape until a fresh covering of snow fills the gaps between them and buries them. The catabatic winds that give rise to very intensive ground winds and low storms form various types of drifts and dunes of relatively small horizontal dimensions (from several metres to several tens of metres), but with steeper slopes than those of the drifts and dunes formed during blizzards. Various shaped ridges are formed as the drifts, dunes and banks harden. There is some dissimilarity in the orientation of the accumulative-deflative forms of the snow surface produced by cyclonic and catabatic winds: the first extend in a latitudinal-wise direction (from east and south-east to west and north-west), and the second in a meridional-wise direction (from south-south-east to north-north-west). A feature of the temperature conditions in this zone is a gradual rise of the mean annual temperature of the air from −50 to −20°C with movement along the slope of the ice sheet towards

the coast. Throughout the year, however, the temperature remains negative and the formation of ice in this, as in the central zone, follows the recrystallization pattern but terminates at a lesser depth because of the considerably greater primary compacting by the wind and the more energetic sublimative recryszallization. In this as in the above-lying zone intensive cooling of the surface continues through the loss of radiation warmth at the expense of very high albedo values (83% at Pionerskaya) and at the expense of long-wave radiation (about 25-30% of the total radiation), as a consequence of which the annual radiation balance is negative (Rusin, 1958). Accumulation of the catabatic wind zone is somewhat less than the quantity of precipitation because part of the precipitation is transported to the coastal zone.

3. The *Antarctic coastal zone* embraces the outer, steeper part of the slope of the ice sheet, measuring from several tens to over 100 kilometres in width. Within this zone the thickness of the ice diminishes sharply through the acceleration of the run-off and the ice drainage differentiates in conformity with the unevenness of the relief below the ice. This is reflected in the isolation of outlet glaciers. The extremely dynamic character of the edge of the ice sheet leads to the formation of a large number of crevasses, pressure banks, hills and ridges. In this zone, which is greatly subject to the influence of cyclones, there is several times more precipitation than in the above-lying zones. The maximum precipitation (600-700 mm) falls at the level where the slope of the ice sheet meets with the lower cloudline (200-1,200 metres above sea level). Insofar as in the summer months the catabatic winds do not reach the coast, dying down several tens of kilometres away from it, precipitation is augmented by a part of the snow transported from the above-lying zone. Because of the violent wind regime the snow cover is distributed very unevenly on the elements of the relief. On mounds and on the bulging bends of the slopes, from where the snow is systematically swept off by winds, evaporation accounts for a considerable expenditure of matter and here ablation frequently exceeds accumulation. In the lower parts of the relief, in the lee of prominent objects, at the coastal barrier and in crevasses snow, on the contrary, accumulates in large quantities. In the coastal zone accumulation in crevasses plays an important role in trapping wind-borne snow. Widening of the crevasses through the movement of the ice sheet leads to the collapse of snow bridges and to the renewal of accumulation in the crevasses. A large quantity of snow accumulates in the narrow coastal strip where catabatic winds abate sharply and deposit the snow they have been carrying from the continent. Enormous snowdrifts form on the leeward side of the barrier, between icebergs, in crevasses and branches of the «deltas» of outlet glaciers, and in the rear part of shelf glaciers. When conditions are favourable, the snowdrifts are evolutionized into wind-produced glaciers. However, snow transported from the continent to the ocean plays a relatively small role in the total balance of the Antarctic ice sheet. The expenditure of matter through the breaking away of icebergs is of decisive importance here. The temperature regime in the coastal zone is not as rigorous as in the interior of the continent. Here the temperature rises to positive values in the summer months and melting plays a part in the process of the transformation of snow and firn into ice (recrystallization-infiltration type of formation of snow). A fohn effect arises when catabatic winds blow, especially in the summer months, as a consequence of which very dry air is observed in the narrow coastal strip and conditions reminiscent of a cold sub-glacial desert take shape in ice-free sections. These sections, known as «oases», have a climate all their own. The relative humidity of the air is here confined to the first few tens of percent, the temperature in the summer months is 4-5°C higher than in the neighbouring sections of the ice sheet, while the surface of the rocks is frequently heated to 20-30°C. In the oases there is practically no stable covering of snow, except for snowdrifts on the leeward side of positive forms of the relief. Evaporation is very intensive. The dry air and the violent winds lead to the formation

of numerous forms of aeolian weathering, while the changes in temperature cause intensive cracking and desquamation of the rock. Chemical weathering takes place in moist and well-heated places. In Antarctica, the «oases» are also the only refuge for life in its most primitive forms. The influence of the melting in the «oases» in the summer months is felt for a distance of several kilometres. This is due both to convective heating and to the pollution of the surface of the neighbouring sections of the ice sheet by the products of the weathering of rocks, which reduce its albedo. Near the «oases», as a consequence of this, snow and ice melt intensively, rivers and lakes form on the ice, and a unique ablative microrelief takes shape on the surface of the glaciers. However, because of the small area in which this phenomenon prevails, it is not of essential importance in the balance of the Antarctic ice sheet.

4. The *extreme ocean zone* occupies the ocean and islands adjoining coast of the Antarctic continent and the outer parts of the shelf glaciers. The boundary between this and the adjoining zone passes along the line where catabatic winds die down from several kilometres to two or three tens of kilometres from the line of the coast. This zone includes the external (northern) parts of the Shackleton, Western and other shelf glaciers, Drigalsky and a number of other islands. A fringe of sea ice and icebergs serves as the outer boundary. The characteristic feature of the ocean zone is the domination of a cyclonic regime, abundant solid and liquid precipitation, extensive cloudiness, frequent fogs, unstable winds, a cold winter and a cool summer. Ice forms along the recrystallization-infiltration and infiltration patterns. In the structure of the layer of snow and firn, strata of firn alternate with infiltration ice. According to observations made on Drigalsky Island and on the Shackleton shelf glacier, the annual precipitation amounts to 700-900 mm.

The process of the ablation of the continental ice (melting of icebergs), which takes a large quantity of heat from the ocean and the air, ends in this zone. The presence of sea ice increases the albedo, on the one hand, and during the melting of this expends heat on the other. The snow, wind-borne from the continent, melts in the same manner. As a result, the climate of this zone is much colder (especially in summer) than might have been expected in these latitudes. The ocean zone is a zone of life in the full sense of the word. As much as the continent itself is dead and deserted, so much is the coastal region surrounding its oceans and islands rich in varieties of life.

Summing up the most important of what has been said above, it may be stated that the central zone of Eastern Antarctica is a region with intensive cooling with little accumulation of precipitation; the catabatic wind zone is a transitional region characterized by a gradual increase of precipitation with distance towards the shore, by an increasing rate of movement of the ice sheet and the transportation by catabatic winds of great masses of cooled air and drift snow from the interior; the coast zone is a region with the maximum precipitation, the maximum rate of movement and the maximum expenditure of the Antarctic ice sheet: in this narrow coastal strip we observe the most intensive exchange of matter and energy (it may be said that the Antarctic ice sheet lives mainly through its edge); the outer oceanic zone is a formation region of shelf glaciers and of marine ices, and also a region of thowing of such ices and of ices brought is from dry land.

The idea that Eastern Antarctic may be divided up into natural zones, which was advanced by the author for the first time after wintering in that area in 1955-1957 (Dolgushin, 1958), has been confirmed and developed by subsequent investigations (Bugayev, 1960).

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THE USE OF STATISTICAL ANALYSIS FOR TRACING OF FIRN LAYERS IN ANTARCTICA

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SUMMARY

The purpose of this research is the compilation of an accumulation map of Antarctica. The identification and tracing of firn layers is part of this process.

An initial report by the present author showed buried firn layers of known date in the Little America region to be traceable over considerable distances due to weather markings, such as crusts from super-cooled rain, for which the history of formation and subsequent metamorphosis had been recorded.

In an effort to increase traceable characteristics statistical analysis of grain size distribution within given layers was attempted. An IBM 650 Computer accomodated 20,000 grain size measurements. Frequency distribution curves were plotted and goodness-of-fit tests applied to determine quantitatively the probability that given samples were from the parent material assumed, ie, the layer being traced. Standard statistical acceptance criteria were adhered to. (Graphs to be presented).

It was determined that sampling error was largely due to a slight sorting of the grains as they spilled on to the measuring plaque. It was possible to minimize laboratory error by having one person make all measurements. The firn layers were found to be traceable, but with notable qualification. On a given traverse, two or three successive pits were often unrecognizable, with good correlation returning thereafter.

When used in conjunction with weather markings the above method is of practical value to the problem of determining accumulation.

A standard practice in geology is to identify sedimentary beds by statistically analyzing the distribution of grain sizes within a given layer. Sediments deposited under the same environment often show a statistically homogenous grain distribution that is characteristic of that sediment over a considerable horizontal distance. This has been the principle applied herein to the Antarctic firn layers.

In a previous report by the present author [1], it was demonstrated that firn layers were traceable over considerable distance by observance of their weather markings, for example, a crust formed by super-cooled rain. Control pits excavated at Little America provided detailed knowledge of the date of formation of a given weather mark phenomenon, and subsequent metamorphosis, for all layers accumulating since the summer of 1956-57. It was shown that stations separated by as much as 600 miles might display identical weather markings.

On the other hand, it was pointed out that the regularity of encountering a given trace marking in a series of traverse pits was not *always* dependable. Thus, investigation was directed toward analysis of the frequency distribution of firn grain sizes as an additional method of tracing layers. Although this paper treats with the specific case of grain size analysis, it should be understood that it is the combination of weather markings and grain size analysis that is recommended for a greater measure of confidence in the problem of layer tracing.

1. MEASUREMENTS

Grain size measurement was made by scraping a 10 × 10 cm tin placque against a snow pit wall and catching the dislodged grains on the placque. The placque had a

[1] Buenos Aires Antarctic Symposium, 1959.

millimeter grid inscribed, against which one could judge grain sizes with the aid of a hand lens. This method of sample collecting is regarded as acceptable in geologic practice [2].

The observers on the Victoria Land Traverse (*) photographed each sample. Thus, the samples have been preserved for detailed laboratory analysis which would not have been practical in the field. In the laboratory, for example, it was possible to compare hand lens measurements against those made more accurately by stereo-comparator. This assisted in determining errors of measurement.

Experiments to isolate measurement error proceeded as follows. One man measured the same sample repeatedly on the stereo-comparator, (8 × magnification, reading to .01 mm). The sample averages differed from one run to the next by approximately 1%, representing the individual's measurement error.

Samples were then divided into sections. Variation between sections occurred from a slight sorting of grains as they spilled on to the placque. The averages of the respective sections of one sample differed up to .11 mm without appearing sorted on the placque. Sorting greater than this became obvious to the eye and bad samples were discarded.

Following the stereo-comparator work, the same samples were read by hand lens as one would do in the field. The hand lens readings differed from the stereo-comparator readings by about 7%.

Lastly, different individuals read the same samples to determine the difference between observers. This was on the order of 16% for the long axes and up to 40% for the short axes.

Also of some consequence, considerable error was found in the markings on the placques. A new scale was substituted when this was discovered.

It was concluded that it was acceptable to make measurements with a hand lens; that all measurements should be made by one person; and that bias was introduced in some cases by slight sorting as grains fell on the placque. Collectively, an error as much as .18 mm could be expected if the observer was the same for all measurements being compared.

2. Discussion of Analysis Method

It has been customary in past geologic problems to compare quartile measures of the frequency distribution curve. This has been dictated virtually by the amount of work involved if one were to use moment measures, the possible alternative (**). Moment measures, however, possess the advantage of increased accuracy. Quartile measures deal with the two center quarters of the frequency curve, leaving the tails of the curve unattended. One can see that in this case all grains of the sample do not contribute to the problem. In some firn samples this can be critical, as there are instances where sublimation has only recently begun, and there are present a scattered 10% of larger crystals which may well be the one characteristic that helps identify that particular layer. Moreover, the range of grain sizes normally encountered with

(2) W. C. Krumbein and F. J. Pettijohn, *Manual of Sedimentary Petrography*, p. 17 (D. Appleton-Century Company, New York, 1938).

(*) The Victoria Land Traverse ranged through considerable environmental change, starting at Little America on the Ross Ice Shelf and extending some 400 miles onto the inland plateau beyond the Victoria Mountains. Because of this variation, and because of the quality and abundance of field data taken by observers, Charles Wilson and Stephen Den Hartog, this traverse was selected for this analysis.

(**) The first quartile is represented by a grain just larger than one-fourth of the distribution, and the second is just larger than three-fourths of the distribution. Moment measures are analogous to physics, the first moment being the center of gravity, the second the radius of gyration.

firn is small, such that one must strive for all the accuracy possible. A modern electronic computer can so assist the investigator as to enable the use of moment measures in approaching this problem.

In the present investigation, more than 20,000 grain size readings were accommodated by an IBM 650 computer. This enabled automatic computation of means, standard deviations, in some instances skewness and kurtosis, the actual plotting of the frequency curve, and application of the various statistical acceptance tests (to be discussed).

The number of measurements in all cases was 47 per sample. This figure was convenient for the computer, as well as reaching into the «large sample» range of the statistical tables.

It was found necessary to experiment with several class interval scales in an effort to have working curves that approached normality. By normality is meant a bell-shaped frequency curve, the items of which are symetrically distributed about the mean. This is because a large segment of routine statistical procedure is based on such a curve. The continuous gradation of the firn grain measurements allows for scale shifting without introducing the bias one would get in the case of discontinuous samples (3).

A conventional arithmetic scale and an adaptation of a log scale, Krumbein's Ø scale, were used as the basis of frequency class intervals.

The value of the arithmetic scale lay in the manifold class intervals used. Measurements were made to .1 mm, and class intervals of .1 mm were used, giving an actual range of .2 mm to 4.7 mm, although grains larger than 2.5 mm were infrequent. This scale preserved the detail distribution of the large-size end of the distribu tion curve. It also graphically illustrated skewness, an identifying characteristic, where use of the Ø scale purposely adjusted such a curve toward normality.

For a full description of the log based Ø scale, the Krumbein reference is suggested. The detail is too involved for the present occasion.

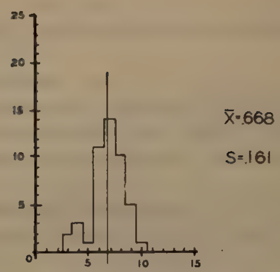
3. Comparison of Histograms

Standard statistical tests were applied to the histograms of all samples in an effort to show which came from statistically homogenous populations (the same parent material). The procedure was as follows.

In order to make those tests which are based on normal distributions, it was first necessary to ascertain whether or not any given curve (or histogram) was acceptably normal. The chi-squared test was first applied. In this test the calculated difference between the curve being tested and a standard normal curve gives a measure of whether or not the sample being tested is sufficiently normal. If the trial curve proves normal (within statistically defined limits) one can apply those statistical tests which are based on assumption of normal a curve.

Where samples were found to be sufficiently normal in their distribution, the statistician's F test was then applied. The F test is essentially a ratio of variances. This ratio should not exceed those ratios obtained from making repeated runs of samples known to be statistically homogenous. For example, a series of samples extracted at one specific location from the same layer of snow would be known to be statistically homogeneous. The discrepancy between the variances of these samples would be attributable to chance, i.e., measurement error and sample extraction inconsistencies. Discrepancy between the above known samples and an unknown sample would have to be within that range in which 95 comparison out of 100 of the

(3) Krumbein and Pettijohn, *op. cit.*, pp. 86-87, for discussion of this choice.



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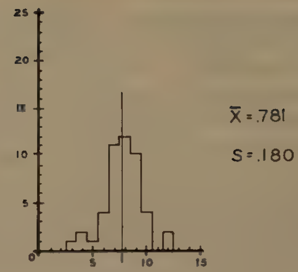

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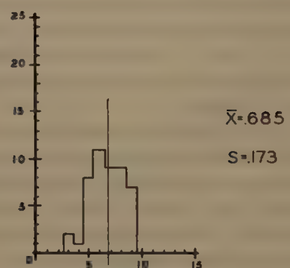

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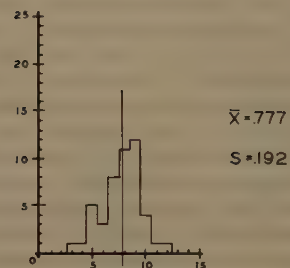

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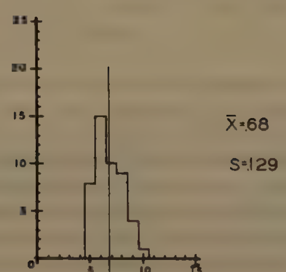

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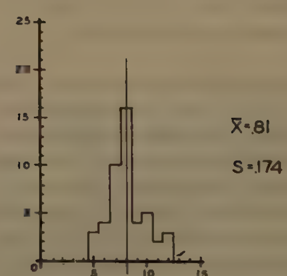

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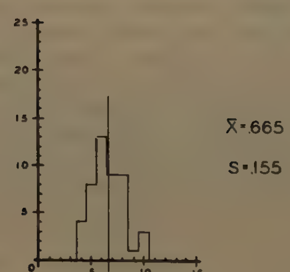

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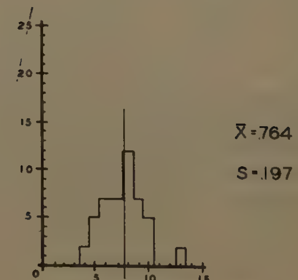

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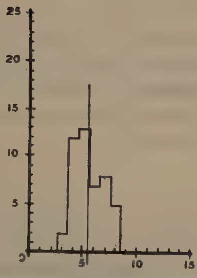

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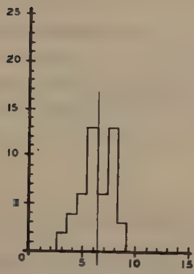

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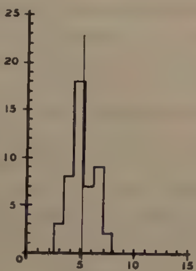

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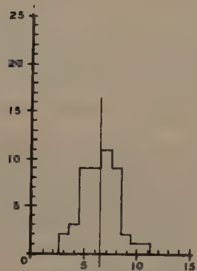

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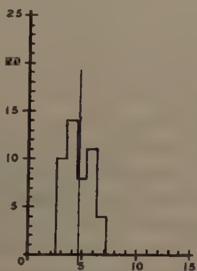

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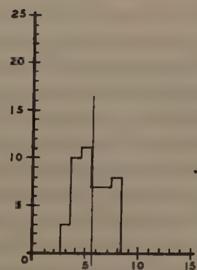

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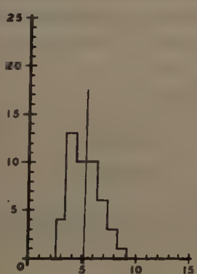
$\bar{X}$=.570

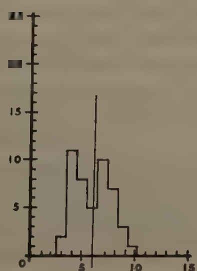
S=.162



81010 I

$\bar{X}$=.530

S=.148



81010 L

$\bar{X}$=.602

S=.180

above known samples would fall. One can view this as the statistician's conventional 5% acceptance level.

Since a normal curve can be completely defined in terms of its variance (or standard deviation squared) and mean, it remains to demonstrate that the means of two curves do not deviate significantly from each other. The Student's t test gives a quantitative evaluation of how the observed difference in two sample means compares with the difference that might logically be expected due to chance.

4. Some Sample Results

Samples were processed in groups of about 20. This was an easily managed quantity and still gave the overlap from layer to layer that helps in spotting change. The layers were first tentatively identified by weather markings, keying them to stratigraphy whose history is known on the adjacent Ross Ice Shelf.

Samples 74127 (R-3a) and 74127 (R-3b) are the same identical grains measured twice by hand lens, thus an indication of the human error. Samples 74127A and 74127B are the same sample as the above, but represent different sections of the sample, thus an indication of the variation within the sample.

The series of samples 67015, 70015, 74035, and 81010 are presumably from the same layer. The layer has been traced over 200 miles with samples extracted every 30 miles. For brevity, four out of seven samples are used for this illustration. The total number of histograms for the entire project number over one thousand (*). In this particular case all the large axes histograms (identification number followed by L) have been plotted in the arithmetic scale. The intermediate axes (identification number followed by I) gave a low percentage of samples whose curves were normal when subjected to a chi-square test against a known normal. The scale was therefore shifted to Krumbein's phi scale in order that the proper probability tests could be applied. The accompanying histograms have means ($\bar{x}$) and standard deviations (S) listed. All have satisfied tests referred to herein.

The histograms thus represent intermediate and large axes of grain populations whose means and variances of size are such that the hypothesis that the samples are from the same parent population has a 95% probability of being correct.

The work referred to in this report had a «salvage rate» of about one in every three samples calculated. This did not necessarily mean that those samples failing the comparison tests were not like samples. It merely meant that many of their distributions could not be conveniently normalized and thus could not be treated in the mass productionmethod for which the computer had been programmed. Rejects also occurred where layers had undergone a high degree of wind sorting, manifested as laminations in the stratigraphy, and in other instances excessive sublimation at depths completely negated use of the data.

When the samples were photographed in the field, the intent at the time did not include the above type of analysis. The present author would like to suggest with the benefit of hindsight, that future work of this nature concentrate on detail. The closer the spacing of sample extraction, the easier the variation within a layer can be recognized. Also the control pit and weather marking analysis mentioned on the first page should be made an integral part of any layer tracing attempt.

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(*) The entire body of this analysis will appear in a report by the Institute of Polar Studies, The Ohio State University.

ANTARCTIC SNOW DRIFT AND MASS TRANSPORT

W.R.J. DINGLE and UWE RADOK

SUMMARY

Relations for the density of drifting snow as function of height and wind velocity are established theoretically and tested by means of observations made at the Australian National Antarctic Research Expeditions base at Wilkes during 1959. The measurements comprise 26 sets of drift snow collections with 8 Mellor type traps at levels ranging from 3 cm to 400 cm and 3 anemometers in winds from 12 m/sec to 30 m/sec at the 10 m level.

The wind observations conform to logarithmic profiles with very small roughness parameters. The drift densities reveal a discontinuity in the particle fall velocity; this occurs between the 25 cm und 50 cm levels and may be connected with the saltation mode of progression prevalent at the lowest levels. The drift density at a given height is an exponential function of the reciprocal of the wind velocity at that height. This gives the appearance of a power law when a restricted velocity range is explored and explains why the exponent in such a law increases with height from less than 0.5 for the 3 cm level to over 5 at 400 cm.

The observed drift snow transport in different layers are exponential functions of the wind velocity at the 10 m level. The total transports between 1 mm and 300 m confirm the very large prelininary estimate previously made for Mawson. However, more than half the total transport is contributed by layers below and above that containing the present measurements. This underlines the need for more information regarding the vertical profile of the katabatic wind.

1. INTRODUCTION

In this paper we present an analysis of a series of 26 drift measurements made by one of us (Dingle) during 1959 at Wilkes Station, using the snow traps described by Mellor (1959) and tested and calibrated in a wind tunnel by Pound (1958); 18 of these runs are complete in every detail. The analysis is made from a physical point of view and compares observed drift characteristics with features predicted by turbulence theory whenever possible. The aim has been to test the operational effectiveness of the drift traps used; this is a prerequisite before any drift transport figures derived from measurements can be accepted as realistic.

2. THEORY

The basis for the theory of steady turbulent snow drift was laid independently by Shiotani and Arai (1953) and Loewe (1956), who showed that the differential equation for the drift density n_z at height z

$$w n_z + (A_z/\varrho)\, \delta n_z/\delta z = 0 \qquad (1)$$

where w is the particle fall velocity, ϱ the air density (assumed constant) and A_z the coeficient of turbulent exchange, has the solution

$$n_z = n_h\, (z/h)^{-w/ku_*} \qquad (2)$$

when the wind profile is logarithmic, i.e. (with the usual symbols)

$$V_z = (u_*/k)\log_e(z/z_0) \qquad (3)$$

and hence $$A_z = ku_x \varrho z \tag{4}$$

as is known to be the case under blizzard conditions (Liljequist (1957)).

When drift density observations are available for a number of levels z_i, $i = 1.2....m$, we can compute the particle fall velocity in the ith layer as

$$w_i^{i+1} = ku_* \log(n_{z_i}/n_{z_{i+1}})/\log(z_{i+1}/z_i) \tag{5}$$

Alternatively comparing each measurement with the logarithmic height average defined by

$$x = 10^{\overset{m}{\Sigma}\log x_i/m} \tag{6}$$

we find

$$\log n_z - \overline{\log n_z} = -(w/ku_*)\ (\log z - \overline{\log z}) \tag{7}$$

a regression equation for log n_z on log z. This form permits the combination of a number of drift runs in order to test the validity of a single value for w for the entire height range investigated, by means of the average deviation of each density n_{z_i} from the regression value $N_{z_i} = \bar{n}_z(z_i/\bar{z})^b$ where b is the regression estimate of the factor $-(w/ku_*)$ in (7).

The dependence of drift density on wind velocity follows from the fact that according to (3) for a given level

$$u_* = V_z k/\log_e(z/z_0) \tag{8}$$

When substituted into the logarithmic form of (2) this gives

$$\log(n_z/n_h) = -(w/k^2)\ \log_e(z/z_0)\log(z/h)\ V_z^{-1} \tag{9}$$

or

$$= -(w/k^2)\ \log_e(h/z_0)\log(z/h)\ V_h^{-1} \tag{9'}$$

Adding these two equations leads to

$$\log n_z - \log n_h = -(w/2k^2)\log(z/h)\,[\log_e(z/z_0)\ V_z^{-1} - \log_e(z_0/h)\ V_h^{-1}] \tag{a0}$$

This indicate that at each level the drift densities should lie on a straight line in a log $n - V^{-1}$ coordinate system, and that the slopes of the lines for two levels, z and h, should differ by

$$\Delta a = (w/2k^2)\log(z/h)\ (\log_e zh - 2\log_e z_0) \tag{11}$$

The relation (11) theoretically determines the drift density—wind relation for al levels when that relation is known for one level either empirically or as function of meteorological factors such as the quantity of loose snow available, its age, etc.

Previous workers have put forward empirical power laws of the form

$$n_z \propto (V_z - B)^\alpha \tag{12}$$

where B is a quasi-constant threshold velocity (cf. e.g. Liljequist (1957) who has $\alpha = 5$, or Stephenson and Lister who imply that $\alpha = 3$). This is not wholly in contradiction with the exponential law $n_z = n(\infty)\ 10^{aV_z^{-1}}$ derived here since for a restricted velocit range the curve $y = a/V$ has only small curvature when plotted against $x = \log V$ as ordinate. It is easy to show that the slope of this curve will be

$$dy/dx = -2.3026\ a/V' \tag{13}$$

where V' is a velocity near the centre of the observed range. Equation (13) suggests that the exponent in (12) should exhibit a height variation in accordance with (11).

We finally determine the drift transport in the layer between the levels z and h. This is

$$Q_z^h = \int_z^h V_z n_z dz = q_h - q_z \text{ where}$$

$$q_z = (1 - w/ku_*)^{-1} z\, n_z \,[V_z - (u_*/k)\,(1 - w/ku_*)^{-1}] \tag{14}$$

for $w/ku_x \neq 1$, whereas for $w/ku_x = 1$

$$q_z = z\, n_z \log_e z \,[V_z - (u_*/2k) \log_e z] \tag{14'}$$

The drift transport Q evidently is a complicated explicit and implicit function of the wind velocity. This does not exclude a simple power or exponential relation, for a restricted velocity range, but whether or not such a relation exists must be determined empirically.

3. Observations

The drift gauging mast at Wilkes Station was located 100 yards to the NE of the former seismic hut which now contained the auxiliary drift gauging equipment. The mast in its final form was devised by the first author after experiments with different arrangements. It consists of a double girder permanently secured to deadmen in the snow and a steel pipe with welded cross arms for 8 drift traps and 3 anemometers. The steel pipe was clamped to the girder by means of 2 U-bolts and could be turned and lowered or raised so as to place the traps and anemometers into the undisturbed air stream at prescribed levels for the majority of blizzard wind directions (070° to 110° true). The proper operating height was achieved simply by letting the lowest of the cross arms just touch the snow surface.

Large rocket-shaped traps were exposed at the 12, 25, 50, 100, 200 and 400 cm levels, while the 12, 6 and 3 cm levels were sampled by small «saltation» traps (cf. Mellor (1959) for trap details). The exposure time in general was 30 minutes for the rocket traps and 2 to 3 minutes for the saltation traps. Each trap was weighed before and after each run by means of laboratory beam balance in the seismic hut. The traps were sealed with wooden plugs before and after exposure. These plugs were removed and replaced in the same order so that every trap was exposed for very nearly the same length of time, although there was a lag of some 20 seconds (depending on conditions) between the exposure periods of successive pairs of traps (400 cm and 200 cm, with 400 cm wind; 100 cm and 50 cm, with 100 cm wind; 25 cm and 12 cm with 25 cm wind). The saltation traps were exposed and weighed during the exposure period of the rocket traps; as a rule two complete saltation runs proved possible in that time. In view of their shorter exposure time the saltation traps gave densities which were not directly comparable with those obtained from the rocket traps. However, one trap of each type operated at the 12 cm level (originally 15 cm; another saltation trap was then used at the 9 cm level), and the ratio of their measurements was used to adjust all the saltation trap figures in each case.

A complete drift gauging run required on the average two hours. On one accasion four separate runs were accomplished in a single day, but more often two runs were made in the same blizzard. The results were transmitted by radio to the Meteorology Department, University of Melbourne, and analysed concurrently throughout the year.

The cup anemometers were periodically exposed together at the same height above the snow surface. Calibration runs in winds between 4 and 55 knots showed their indications to be identical for practical purposes.

4. Results

4.1. *Wind characteristics*

The winds measured at 3 levels on the drift gauging mast throughout conformed to logarithmic profiles. The values of the roughness length z_0 ranged from 2×10^{-1} cm to 10^{-6} cm and showed no dependence on wind speed. However, z_0 seemed to increase slightly with the drift density, and the largest roughness values occurred in the presence of patches of soft drift on the more usual hard-packed snow surface. An example of this effect is shown in table 1:

TABLE 1

Roughness lengths and surfac conditions for September 29th, 1959

time (z)	0100	0300	0630	0800
z_0 mm	4.6×10^{-2}	7.4×10^{-2}	4.1×10^{-3}	5.8×10^{-3}
V_{10} m/sec	29.9	24.3	25.2	23.6
surface conditions	Surface snow generally hard packed with a lot of soft patches wich have formed drifts on top of the hard packed snow		Tendency for loose drifts to disappear	Surface snow hard packed; most of the loose stuff has disappeared

The (logarithmic) mean value of z_0 for 9 clear-cut cases of bare hard-packed snow was 1.8×10^{-3} mm, as against 5.6×10^{-2} mm for 6 cases of soft snow patches. Both values are well below the roughness lenths found by Liljequist at Maudheim; this may possibly be characteristic of katabatic wind conditions.

For a test of some of the theoretical results of section 2 a representative z_0 value for the present data is required. This can be found in different ways. Mean relative velocities V_z/V_{100} (Sverdrup 1936) gave $\bar{z}_0 = 8.5 \times 10^{-3}$ mm for all 26 runs and 6.3×10^{-3} mm for the 18 runs with measurable drift amounts at all levels. A different estimate can be based on the shear velocities u_* and the observed winds at one level by interpreting (8) as a regression equation of the form

$$u_* = \beta V_z; \text{ since on the other hand}$$
$$\beta = k/\log_e z/z_0$$

we obtain

$$z_0 = ze^{-k/\beta} \tag{15}$$

Equation (15) when applied to the 400, 100 and 25 cm levels yields values for z_0 somewhat below those derived from relative winds, viz. 1.6×10^{-3} mm, 3.7×10^{-3} mm, and 4.6×10^{-5} mm. The largest of these will be used below as representative for the 18 complete drift runs.

4.2. *Drift density as function of height*

In evaluating drift densities a constant trap efficiency of 0.74 has been assumed. While in the wind tunnel the efficiency tended to increase very slightly with wind velocity (Pound 1958) this may well be offset under operational conditions by greater snow losses through the trap exit.

Fig. 1 shows 3 representative examples of drift density profiles, spanning a range from almost the equivalent of the air density to less than 10^{-4} times that quantity. Two of the profiles approximate to straight lines while in the third there is a suggestion of curvature, or perhaps, of a discontinuity of slope. That the latter is, in fact, the proper interpretation is shown by the average logarithmic density deviations from the regression estimates for different levels, plotted in fig. 2a (crosses). Evidently the average fall velocity underwent an abrupt change between 25 and 50 cm but was

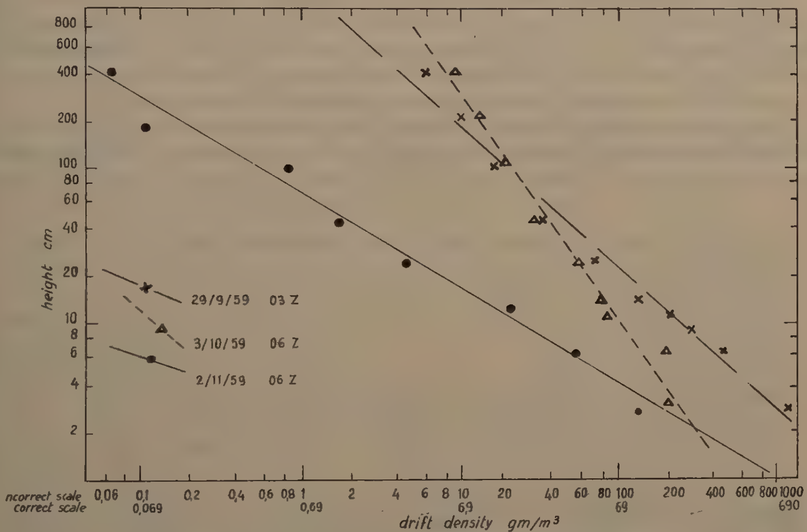

Fig. 1 — Observed drift density profiles.

Note that in accord with the footnote, the horizontal scale must be displaced to the right to bring the values 6,9, 69, etc. to the positions of 10, 100, etc. of the existing scale.

substantially constant at higher and lower levels (as shown by the triangles in fig. 2a which represent deviations from two separate regression lines for the ranges 3-25 cm and 50-400 cm). This is confirmed also by the fall velocities for individual layers, computed by means of (5); these were found to be independent of wind velocity and their averages are shown in fig. 2b. The change in fall velocity may be related to the «saltation» mode of progression; calculations now in progress (cf. Mellor and Radok 1959) show that the greatest heights attainable by a bouncing snow particle in a 30 m/sec wind (at the 10 m level) is of the order of 30 cm. But whatever its reason

(*) By a regrettable error the trap efficiency has been applied twice in the calculations. This has the effect of rendering all drift densities and amounts too large by a factor of $0.69^{-1} = 1.45$, but does not affect their relative magnitudes which are correct.

the discontinuity in the fall velocity makes it advisable to establish separate drift density-height profiles for the layers below and above 50 cm, especially when extrapolations to heights beyond the observed range are to be made.

4.3. *Drift density as function of wind velocity*

Following the argument in section 2, regression equations were computed for the logarithmic drift density at each level as function of the reciprocal of the wind velocity at that level. These equations are summarized in table 2. The quantity S represents the proportion of the variation in log n_z explained by the regression; S is related to the standard deviation σ_a of the regression coefficient a by

$$(\sigma_a/a)^2 = (S^{-1}-1)/p - 2 \qquad (16)$$

where p is the number of observations at each level (here 18).

Fig. 2c show the slopes of the log n_z — V_z^{-1} regression lines plotted against height (full curve). The theoretical estimates derived from equation (13) for the observed 400 cm regression, $z_0 = 3.7 \times 10^{-3}$ mm, and a mean fall velocity of 25 cm/sec are shown by the broken line. The theoretical height variation in the drift density—wind relation appears to be well reproduced by the present drift observations.

The same drift densities have also been used for regression estimates of the exponents α in equation (12) which are included in table 2 and plotted in fig. 2d. The exponent increases from less than 0.5 at 3 cm to just over 5, Liljequist's value, at 400 cm

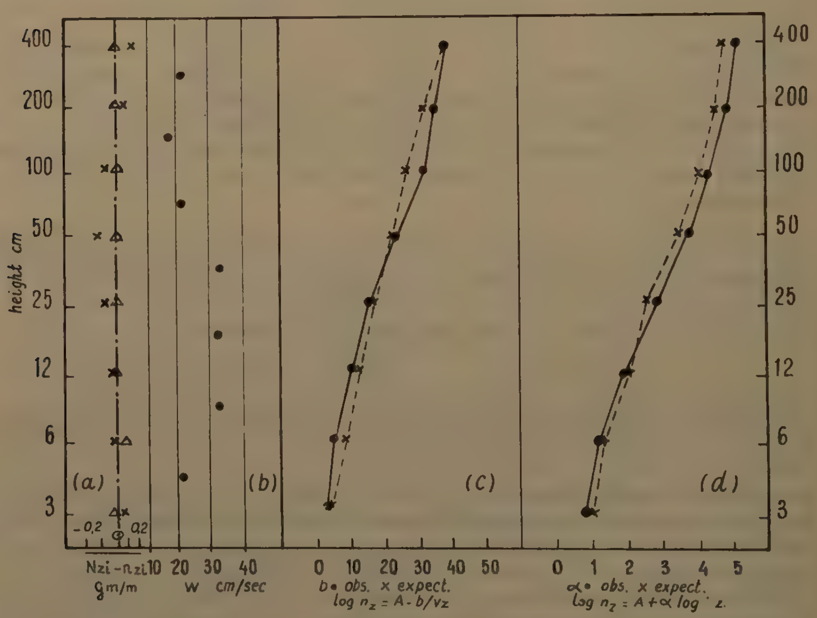

Fig. 2 — Drift parameters as function of height.
a) Deviations from the profile of equation (4) with single constant w (crosses) and with different w for the layers 3-25 cm and 50-400 cm (triangles).
b) mean fall velocities, cm/sec.
c) observed and expected regression coefficients for given 400 cm regression of log n_n on v_n^{-1}.
d) observed and expected exponents in power law (12).

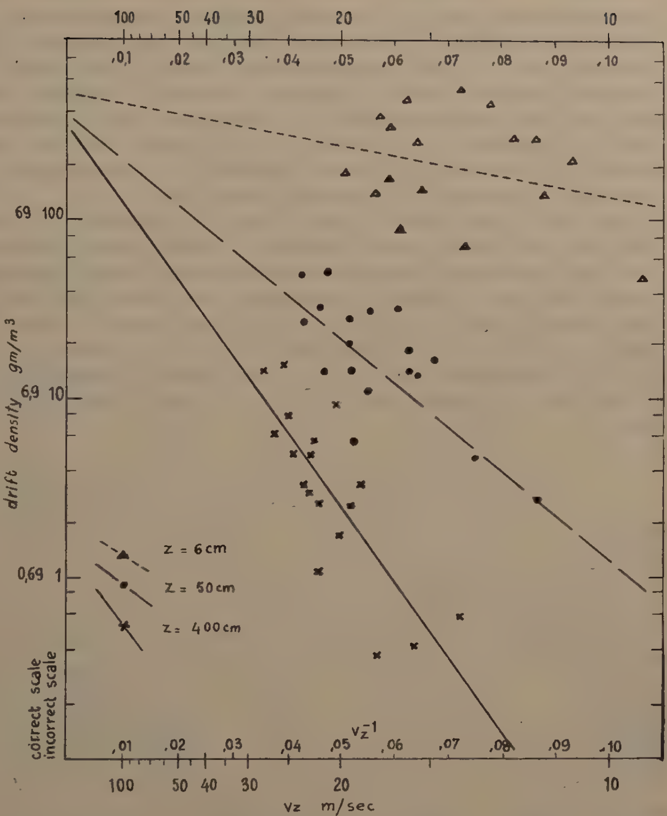

Fig. 3 — Observed drift densities at 3 levels as function of reciprocal wind velocity.

Note that in accord with the footnote, the horizontal scale must be displaced to the right to bring the values 6,9, 69, etc. to the positions of 10, 100, etc. of the existing scale.

However, using (13) with the centre of the observed velocity range as V′ in each case very similar exponents are obtained (fig. 2d, broken line). The S and S' figures in table 3 show that the fit of the V^{-1} regression is slightly better than that of the log V regression at all but one level. The remainder of the variation represents partly meteorological factors and for the rest errors of observation; these are evidently rather large at the lowest levels.

Under the power law the drift density increases indefinitely as $V \to \infty$, whereas under the exponential law it approches a finite value $n(\infty)$. Table 3 suggests that $n(\infty)$ has the same order of magnitude for every level. This is further illustrated in fig. 3 which suggests for very high wind velocities a uniform mean drift density of around 500 gm/m³ over the height range here studied.

4.4. *Drift transport*

Drift transport figures have been computed for each of the 18 complete runs and the following layers: 1 mm-3 cm; 3-50 cm; 50-400 cm; 4-100 m; and 100-300 m. As it turned out each of these layers contributed roughly in the same measure to the

total drift transport. For the highest layers a linearly decreasing exchange coefficient was assumed, and the total mass of drift snow in the layer was multiplied by a mean velocity, arbitrarily assumed as 5/6 of the maximum wind, corresponding to the logarithmic wind profile and $z = 100$ m; this is essentially Loewe's (1956) procedure.

The computed transport values all suggested exponential functions of the 10 m wind velocity. The corresponding regression equations for $\log Q$ on V_{10} are summarized in table 3. The total transport values are shown in fig. 4 together with their regression line (which accounts for over 80% of the total variation) and its extreme slopes (3 standard deviation limits compatible with sampling fluctuations.

A full discussion of the transport figures and their implications is beyond the scope of this paper, and we must restrict ourselves to a comparison between the present estimates (*) on the one hand and two earlier ones for Mawson and Port Martin on the other. Allowing for slightly different assumptions at the higher levels the present estimate for a 28 m/sec wind, 136 gm/cm sec, is practically identical with the preliminary estimate for a 28 m/sec wind at Mawson, 153 gm/cm sec (Mellor and Radok 1959. For a 36 m/sec wind at Port Martin a purposely conservative estimate by Loewe (1956) was 28 gm/cm sec whereas the present data suggest 840 gm/cm sec — 30 times as much.

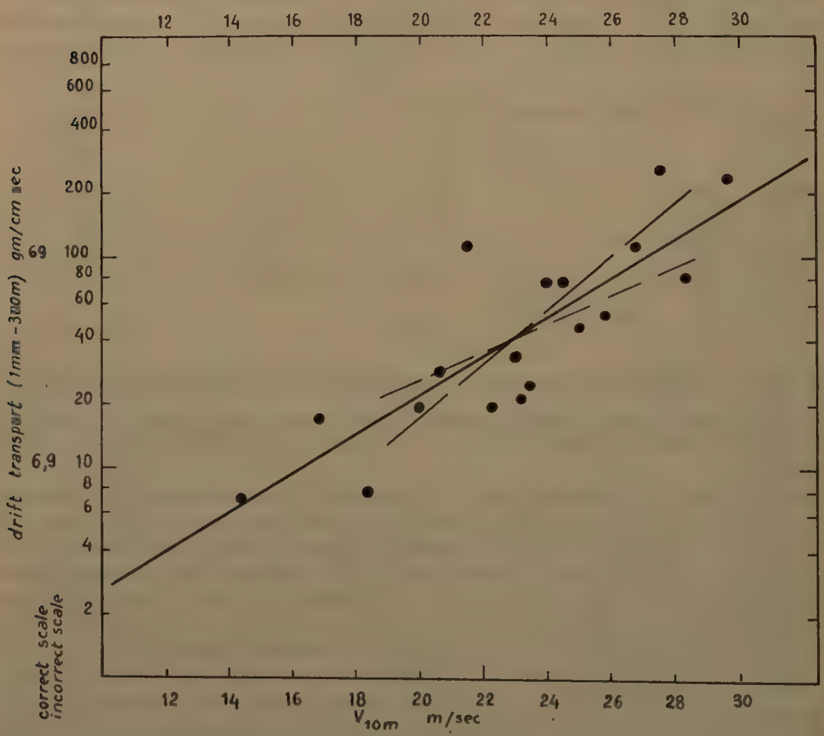

Fig. 4 — Drifts transport between lmm and 300 m as function of wind velocity at the 10 m level.

Note that in accord with the footnote, the horizontal scale must be displaced to the right to bring the values 6,9, 69, etc. to the positions of 10, 100, etc. of the existing scale.

TABLE 2

Regression constants for log drift density as function of reciprocal and log wind velocity

$$\log n_z = A + aV_z^{-1} \text{ or } n_z = n_z(\infty)\, 10^{aV_z^{-1}}$$

$$\log n'_z = A' + \alpha \log V_z \text{ or } n'_z = n'_z(0)\, V_z^{\alpha}$$

height cm	3	6	12	25	50	100	200	400
a m/sec	− 3.22	− 5.50	− 11.12	− 17.63	− 25.08	− 35.73	− 38.39	− 42.26
A	2.808	2.715	2.688	2.640	2.621	2.877	2.705	2.537
$n(\infty)$gm/m³	643	519	488	436	417	753	508	344
$S = \dfrac{a^2 - \text{var } V_z^{-1}}{\text{var log } n_z}$ %	3.4	8.5	27.8	43.6	45.2	67.6	68.1	64.6
α	0.457	0.803	1.624	2.471	3.525	4.173	4.800	5.120
A'	1.978	1.395	0.024	− 1.462	− 3.215	− 4.374	− 5.491	− 6.279
$n'_z(0)$ gm/m³	95.1	24.8	1.06	3.45	6.01	4.23	323	5.26
$S' = \dfrac{\alpha^2 \text{ var log } V_z}{\text{var log } n'_z}$ %	2.0	6.2	23.5	40.5	44.8	52.4	68.0	66.8

Note: var $x = \Sigma(x - \bar{x})^2/17$; S is proportion of var log n_z explained by the regression.

(*) In accord with the footnote p. 5 the values A, A″, and C, should be reduced by log 1.45 = 0.162, while n_z, n_z (O), and $Q(0)$, must be divided by 1.45.

TABLE 3

Regression constants for log drift transport in different layers as function of wind velocity at 10 m.

$$\log Q_{z1}^{z2} = C + c V_{10} \text{ or } Q_{z1}^{z2} = Q(0)\, 10^{c V_{10}}$$

Layer	1 mm-3cm	3-50 cm	50-400 cm	4-100 m	100-300 m	1 mm-300 m
c sec/m	0.0449	0.0568	0.0936	0.1578	0.1661	0.0974
C	1.716	1.388	0.521	−0.483	− 0.151	1.095
$Q(0)$ gm/m sec	52.0	24.4	3.32	0.329	0.706	26.0
$S = \frac{c^2 \text{ var } V_{10}}{\text{var } \log Q}$ %	13.0	42.9	58.8	74.0	72.5	83.5

It has been shown that measurements with the drift traps of Mellor (1959) have provided the expected features for some drift characteristics which can be predicted by turbulence theory. Furthermore the agreement achieved in the drift transport estimates for two different places on the Antarctic coast is encouraging. However, these estimates evidently remain in doubt without detailed information about the vertical profiles of the katabatic wind speed and direction throughout the layer containing drift snow. Attempts are being made this year at Wilkes to obtain this information. A further point in need of study is the local variation of drift intensity, both at the coast and in the interior of Antarctica, since this must somehow render the available amount of snow compatible with the enormous quantities involved in snow drift.

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(*) The present estimates should read 94 and 580 gm/cm sec. (instead of 136 and 840 gm/cem sec). This suggests more intense drift at Mawson which is closer to the continental ice slope than Wilkes.

RESULTS A OF STUDY OF THE PROCESSES OF FORMATION AND STRUCTURE OF THE UPPER LAYER OF THE ICE SHEET IN EASTERN ANTARCTICA

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SUMMARY

The principal snow accumulation in the outside regions of the mainland takes place during the cyclonic weather, accompanied by strong E and ESE winds, intensive snowfalls and comparatively high temperature. Catabatic S and SE winds are almost not involved in the accumulation of snow, more often they destroy that was deposited during the eastern winds. The snow accumulation is characterized by its irregularity. Under the action of wind there arises a complicated microrelief of the snow surface consisting of accumulative (snow-drifts, dunes, barkhans) and erosive (sastrugi) forms. The very existence and the size of sastrugi depends on the strength of catabatic winds. The greatest sastrugi, therefore, are in the zone of catabatic winds.

The density of freshly deposited snow depends on the strength of the wind and the temperature of the air. In consequence of this gradual decrease of the density of snow from the sea coast (0,45) towards the center of the mainland (0,30) is observed except for the belt of maximum of catabatic winds (0,48). The density on the ice shells decreases as the latitude increases.

The intensive thawing of the snow cover takes place only up to the height about 400 m. above the sea level; the weak thawing occurs up to the height 1000 m. In the central regions of the mainland in summer there occurs only the radiation melting of crystalls leading to the formation of radiation crusts. Depending on the intensivity of the thawing in Antarctica four zones of the ice formation are distinguished : infiltration - congelation (up to 1,5 km from the coast), cold infiltration (up to 10-13 km), infiltration - recrystallisation (up to 50-100 km) and recrystallisation ones.

In snow-firn cover two groups of the snow layers — «winter» and «summer» ones — are sharply marked. The winter layers are characterized by : a) large hardness, b) large density and consequently the smaller porosity, c) the small size of crystalls, d) presence of wind crusts. The summer layers on the contrary are characterized by : a) small hardness, b) small density and consequently the greater porosity, c) the comparatively large size of crystalls, d) the presence of radiation crusts and strata of infiltration firn and ice. These indications are the reliable criterion for distinction of the seasonal snow layers and they allow to determine the age of the upper part of the ice sheet with a high exactness.

The processes of the formation and structure of the upper layers of the ice sheet were studied in detail in 1956-1958 by the Soviet Antarctic Expedition headed by P.A. Shumsky. Physical and geographical methods of investigation were used. In conformity with the tasks that had been set, all the investigations were divided into stationary, route and laboratory.

Stationary work was done in the region of Mirny and six kilometres from the coast. The following was done to study the formation of the snow sheet: a) crystallographic study of solid precipitation during snowfalls and snowstorms; b) blizzard measurements; c) snow-measurement observations with rods and cables stretched over the surface; d) observations of the processes of the fall, transportation, accumulation and blowing away of snow and other forms of precipitation, the formation, movement and destruction of various forms of the microrelief of the surface of the sheet, the appearance of crusts of various origin, evaporation, melting and infiltration, and so forth with the recording of the results of these processes showing the changes in the height of the surface, its microrelief, density and hardness of the surface layer.

There were specially equipped plots of ground measuring 50 $\times$ 50 metres. Plot No. 1 was situated 600 metres away from Mirny on the side of a gently sloping hollow

in the cold infiltration (firn) zone of the formation of ice. Plot no. 2 was situated six kilometres away from the coast (165 metres above sea level) on a flat terrace-like ledge, that was characteristic for the lower 10 kilometre-wide coastal strip of the continent, in the very heart of the firn zone.

In the conditions that obtained on the surface of the ice—huge flat expanses, predominance of strong winds and, as a consequence of this, considerable unsteadiness of the microrelief and an immense difference of its form in the higher altitudes—rod measurements, particularly single measurements, do not give a correct picture of the dynamics. Consequently, we took the measures with the aid of cables, which were stretched in two directions (in the direction of the dominant winds and across) and marked with knots at every metre. The height of the cable above the surface of the snow was measured at every knot.

In addition to measuring the thickness of the snow cover with the help of cables, the different forms of the microrelief were photographed regularly, the surface of the snow was recorded and sketched to show the age and structure of the different sections. A comparison of the drawings of these plots clearly shows the development of the surface of the ice sheet. The hardness of the surface was measured with a hardness indicator gauge. The density was measured by weighing samples of snow in a cold laboratory on very accurate technical scales. A series of blizzard measurements were taken during characteristic types of weather.

The study of the development of the thickness of the snow in time was studied in the region of Mirny, where in a plot measuring 5 $\times$ 5 metres the surface was marked at different times with cord, pits were dug monthly to cover the whole depth of the seasonal layer and monoliths of snow were selected for laboratory investigation. The process of sinking was studied at Pionerskaya, at a point six kilometres away from the coast and in the region of Mirny with the aid of sinking self-recorders designed for the continuous registration of the sinking of firn.

Route investigations embraced the entire region covered by the Soviet expedition. Fifty stakes from which readings of the height of the snow sheet were taken six times in the course of the year were driven into the ice along a 50-kilometre radial profile at Mirny where the non-differential edge of the ice sheet with a narrow zone of melting at the coast is typical of the coast of Eastern Antarctica. More detailed investigations were conducted in a section up to six kilometres wide in the coastal strip, where the absolute height increases rapidly and the conditions of the relief change sharply. Here stakes were driven into the snow at intervals of 200 metres and readings taken twice a month on the average.

Field glaciological investigations were conducted by parties that came to the investigated spot by aircraft or tracked vehicle. These investigations were conducted on three occasions on the surface of the Shackleton ice shelf, on Mill, Bowman and Masson ice domes, and at the inland bases Vostok-1 and Komsomolskaya. Detailed investigations embraced Drigalsky Island, where a network of 38 stakes was disposed.

The field glaciological investigations, which became standard during the process of the work, included: a) digging 2-3-metre pits at intervals of up to six metres; b) selecting samples of snow from the entire depth of the pits and subjecting them to all-round laboratory investigation; c) measuring the hardness of the snow at a depth of from four to eight metres with a hardness-gauge-sound; d) recording the microrelief of the surface of the snow with snow of different age and determining its surface hardness, density and the direction of the predominant forms.

The structure of the snow-firn sheet and the properties of the snow were studied in the laboratory. For this purpose segments were cut from snow monoliths measuring 50 $\times$ 30 $\times$ 40 cm, examined and photographed; layers of different structure were distinguished. The following was determined to distinguish the layers in the laboratory: density—porosity, airtightness in the vertical and horizontal directions, height of

the capillary uplight, maximum water-holding capacity and structure, i.e., form, dimensions, reciprocal arrangement and orientation of crystals, as well as the form and dimensions of the air inclusions, microphotography.

The snow sheet is formed chiefly by snow falling in these areas and only to a very small extent by snow carried by winds from the interior of the continent. In the coastal strip up to 600 kilometres wide the chief sources of nourishment of the glacier are precipitation during cyclones with strong easterly winds, and intensive snowfall at a relatively high temperature (from −5 to −15°C.). In the central regions of the continent, on the coastal glacial domes and on the surface of the ice shelf an important role is played by hoar-frost that forms on the cold surface of the snow.

There is a direct connection between the fall of solid precipitation and temperature. Along the coast the precipitation in early spring and late autumn consists chiefly of crystals of a lamellar type of growth, which are typical for temperatures below −23°C. Pillar-shaped crystals are predominant throughout the year in the interior of the continent, and along the coast in winter, but lamellar-shaped crystals have been frequently observed during deep cyclones even as far inland as Pionerskaya. Around Mirny the average for a year consists of 75% lamellar-shaped crystals and 25% pillar-like crystals, while at Pionerskaya there are only 45% lamellar-shaped crystals and 55% pillar-like crystals. P. A. Shumsky (1957) has observed a considerable quantity of pillar-like forms 125 kilometres from the coast at an elevation of 1,600 metres.

Two periods are distinguished in the formation of the snow sheet: cyclonic, which yields material for accumulation, and antycyclonic, which redistributes this material. The intensity of the accumulation of snow during cyclonic weather depends on the intensity of the snowfall, the wind velocity and the temperature of the air. The accumulation of snow is directly proportional to the intensity of the snowfall, and decreases when the wind velocity rises, all other conditions remaining the same. The snow sheet grows at a very rapid rate during cyclones: along the coast in autumn up to 6-7 cm a day, and in spring up to 3-5 cm a day. The weak winter cyclones add little to the snow sheet.

Strong south-easterly catabatic winds usually destroy what has been accumulated during easterly winds. In the course of 1957, the surface in the region of Mirny dropped by 12 cm through the action of catabatic winds. Accumulation or dispersal during catabatic winds is practically independent of temperature, the chief reason for these processes being the wind velocity. When the wind velocity is above 14-16 metres per second, destruction begins to be predominant over accumulation. This phenomenon is explained by the mechanism of drifting snow.

Over the larger part of the continent the snow sheet forms under the action of winds. Blizzards and lower storms are well developed. Blizzards breaking out during cyclonic weather are usually accompanied by easterly and east-south-easterly winds. Lower storms are usually accompanied by catabatic south-easterly and south-south-easterly winds. During storms a direct dependence of the intensity of the drift (i) on wind velocity (V) is observed. This is expressed by the formula $i = 0{,}41V - 2.5$ (Fig. 1). i is equal to zero during a wind velocity of 6.1 metres per second (*). This means that a horizontal displacement of snow particles begins during a wind velocty of about 6 metres per second, while under a slower velocity there is simply a snowfall which is not accompanied by drifts.

This direct dependence of the intensity of the drift on wind velocity during blizzadrs (as distinct from low storms) is due to the fact that during blizzards masses of snow enter the snow-wind current from above. But with the given wind velocity only a limited quantity of snow may be retained in the air, the surplus, which comes

(*) Here and further the wind velocity is given for a height of 10 metres above the surface.

from above, leading to the accumulation of a similar quantity of snow on the surface. Incidentally, this circumstance is the reason for the more or less uniform accumulation of snow during blizzards despite even high wind velocities. The fact that masses of snow come from above and not from below makes the intensity less dependent on the condition of the bedding surface.

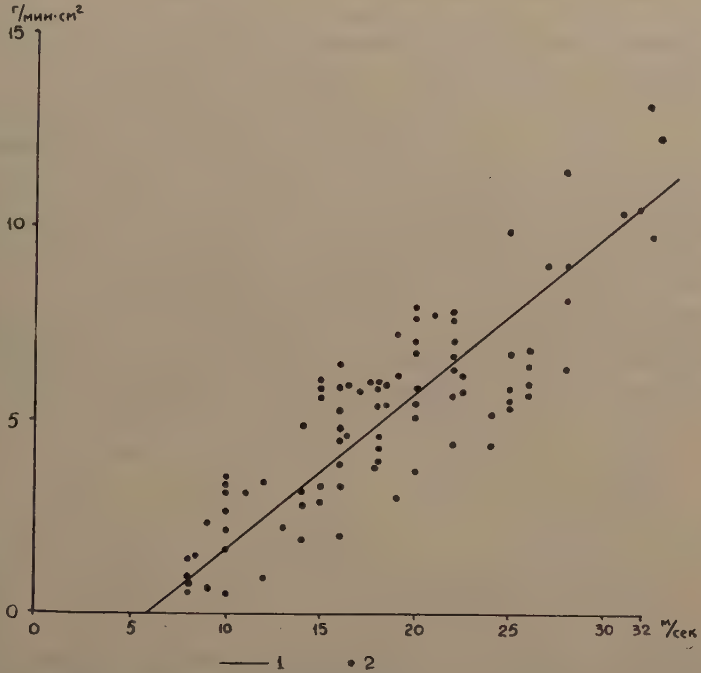

Fig. 1 — Dependence of the intensity of drifting on wind velocity during a blizzard $i = 0.41V - 2.5$.

Snow drifts in a totally different way during low storms. In such cases the higher intensity of the storms leads to a sharp reaccumulation of snow, and the greater the wind velocity is the more uneven will this accumulation be. The intensity of the drift during low storms depends, in addition to wind velocity, mainly on the condition of the bedding surface. Fig. 2 gives the results of measurements of the intensity of low storms under different surface conditions: loose (hardness less than 1 kilogram per square cm), medium (1-5 kilograms per square cm) and hard (more than 5 kilograms per square cm). The minimum wind velocity under which the snow begins to drift, depends on the hardness of the surface. This velocity ranges from 5 to 10 metres per second, a higher velocity being required when the snow is compact.

A rise of wind velocity at a definite moment leads to a sharp leap in the intensity of the storm. This leap means that a strong ground wind has developed into a low storm. At the moment a ground wind develops into a storm, an increase in wind velocity by 1-2 metres per second causes the intensity of the drift to rise sometimes by 4-5 grams per square cm per minute. Where the surface is hard, this leap takes place during a wind velocity of 16.5 metres per second, and where there is a medium surface

—during 15 metres per second, and a loose surface—14.5 metres per second. This is the very circumstance, which with catabatic winds of more than 14-16 metres per second always cause the destruction of the surface, at the expense of which additional masses of snow enter the snow-wind current.

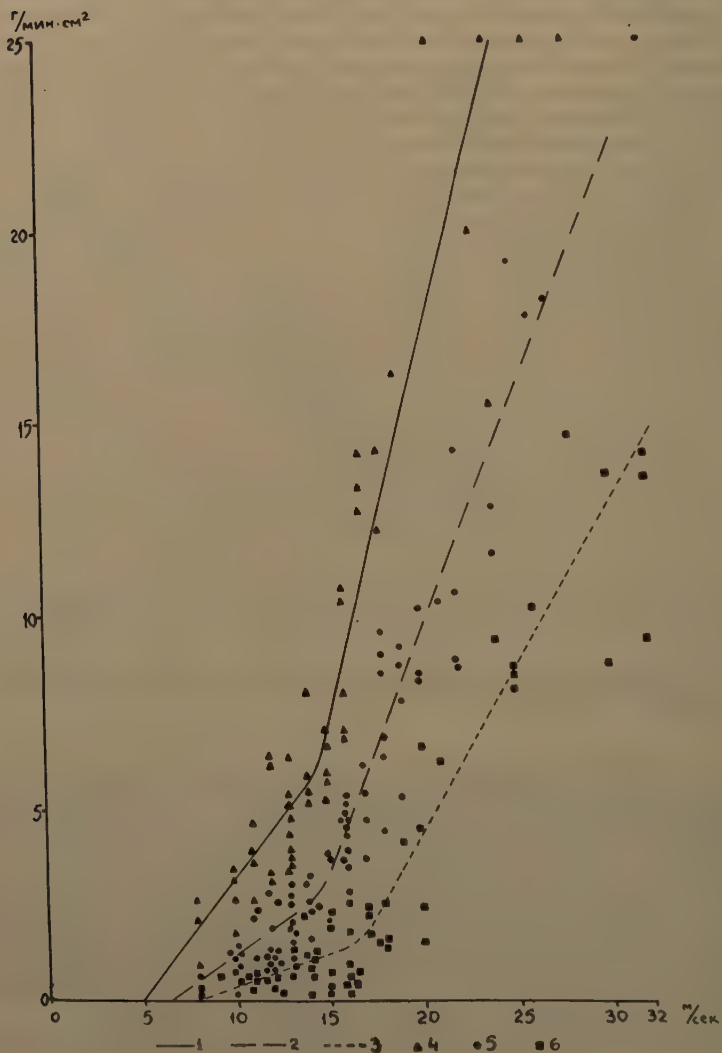

Fig. 2 — Dependence of the intensity of drifting on wind velocity during low storms. Conventional symbols:
1 and 4 — predominance of loose snow on the surface;
2 and 5 — predominance of snow of medium hardness on the surface;
3 and 6 — predominance of hard snow on the surface.

Rare cases of the accumulation of snow accompanied by weak winds lead to the formation of loose fresh snow, whose density fluctuates from 0.10 to 0.25 grams per cubic cm, and its hardness is only 0.04-0.15 kilogram per square cm. The result of this insignificant hardness is that even a wind with a velocity of 5-7 metres per second begins to displace the snow. That is why such snow usually exists for only about a day.

As the wind velocity rises, the density of the fresh snow increases. In the temperature range from $-5°C$. to $-10°C$., during which the main bulk of fresh snow accumulates, its density (γ) is parabolically dependent on the wind velocity (V). (Fig. 3). This dependence is expressed by the formula $\gamma = 0.104 \sqrt{V - 6}$. Under a wind velocity of less than 6 metres per second, the density of the accumulated snow is close to zero, the explanation for this being the absence during low-velocity winds of a compacting action by the transported particles of snow.

The forms of the snow sheet rising as a result of the action of winds are: free accumulative, forced accumulative and erosion. Free accumulative forms take shape on considerable even expanses. They are either loose snowdrifts or flat gently sloping bulging mounds, or steeper and bigger snowdrifts. Large ridges occurring over distances of from 10 to 400 kilometres are characteristic Antarctic formations. Forced accumulative forms take shape near big protuberances or on uneven ground. The development of moving snowdrifts is a characteristic feature in the coastal strip of the continent during anticyclonic weather in winter. Strong (particularly catabatic) winds destroy the surface formations, lay bare deeper layers, giving shape to two main erosion

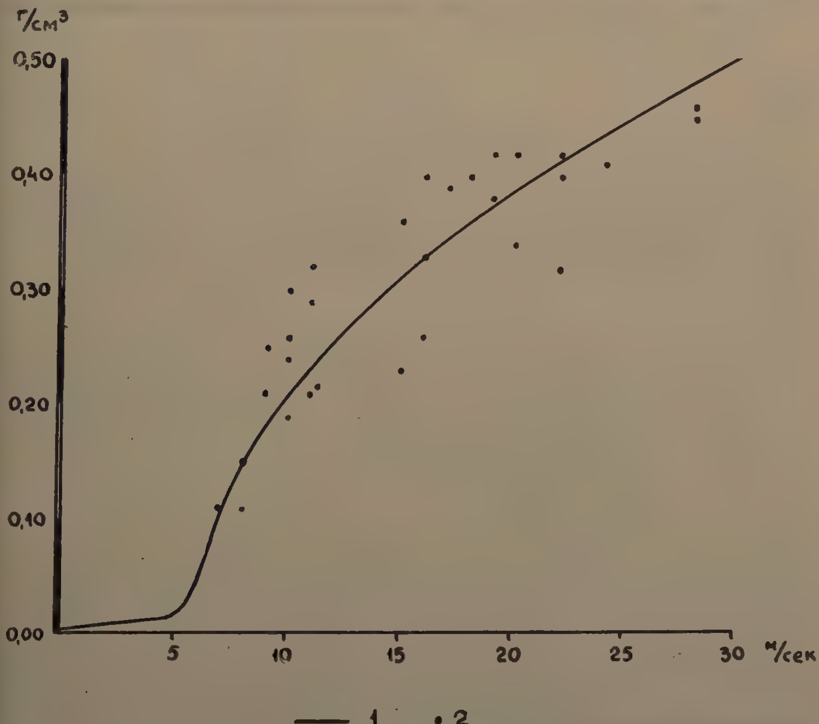

Fig. 3 — Dependence of the density of freshly deposited snow on wind velocity. $\gamma = 0.104 \sqrt{V - 6}$.

forms: sastrugi and hollows. Sastrugi and fields of sastrugi are the main feature of the microrelief of the Antarctic surface in the zone of catabatic winds.

The intensity of the destruction of the snow sheet mostly depends on wind velocity and the hardness of the snow surface. The maximum hardness of the snow (H) was found to be directly dependent on the velocity (V) of the winds destroying it in the range of temperatures (from —12°C. to —15°C.) prevalent along the coast of the continent in winter. This is expressed by the formula $H = 0.15\ V - 0.75$ (Fig. 4). No matter how loose it is, the surface is not destroyed when wind velocity falls below 5 metres per second. This figure conforms well with the data obtained for the drifting of snow, where the value 5 metres per second is the minimum wind velocity under which snow begins to be transported during a ground wind.

The law-governed character of the changes of the natural conditions in Antarctica farther away from the coast gives rise to a regular distribution of the density of the upper layer of the ice-sheet over the territory of the continent. Wind velocity is the main force influencing the density of the snow. In the coastal regions of the continent and on the ice shelfs the considerable density is due to the great force of the cyclonic winds, and in the strip from 100 to 400 kilometres this is due to the maxium development of the catabatic winds. Along the radial profile from Mirny to the Pole of Relative Inaccessibility the density varies as follows: Shackleton ice shelf—0.45 grams per cubic cm; coastal strip of the continent—0.41-0.45; region 225-275 kilometres—0.46-0.48; Pionerskaya Base—0.41; Vostok-1—0.39; Komsomolskaya—0.37; Sovetskaya—0.31; Pole of Relative Inaccessibility—0.32.

The law governing the distribution of sastrugi is analogous. In winter the main direction of the sastrugi in the strip of the continent right up to 900 kilometres from the

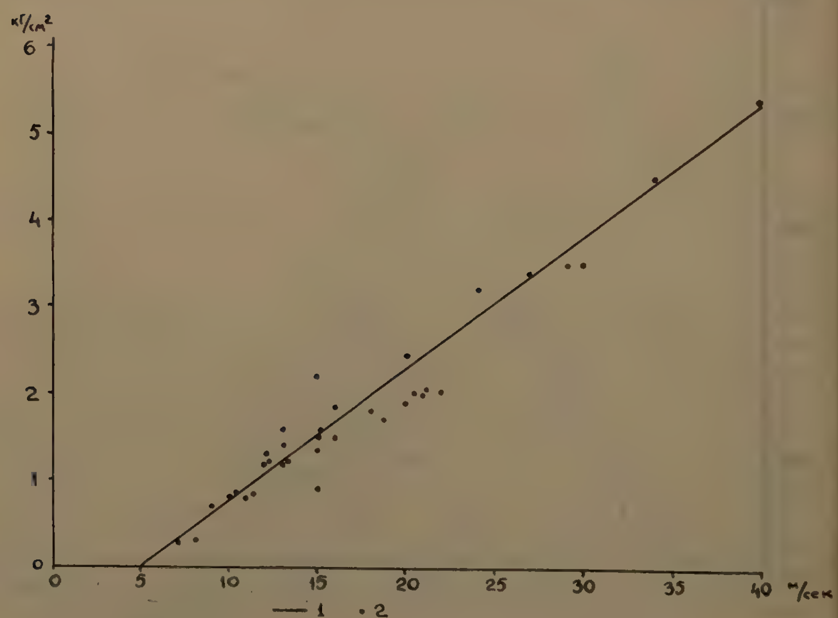

Fig. 4 — Dependence of maximum hard snow during its destruction by wind of a definite force. $H = 0.15\ V - 0.75$.

shore is south-east (azimuth 200-240°) and east (240-280°). Small forms of microrelief, not exceeding 30-40 cm, are a feature of the coastal zones of the continent. The formation of tall sastrugi is hindered by the abundant accumulation of snow covering them. The biggest sastrugi, up to 1.5 metres high, occur in the region from 225 to 275 kilometres from the coast. Farther south they level out, and in the region of Komsomolskaya height does not exceed 25 cm, while at Vostok Base their maximum height is 30-40 cm.

The upper part of the snow-firn consists of layers of fresh snow which were formed during the winter and are distinguished from the lower layers by the fact that they excellently retain the primary structures and textural features called forth by the accumulation of snow. Snow accumulates in many layers, a feature which reflects wind velocity changes within a wide range. The layers of fresh snow consist of crystals measuring only 0.2-0.3 mm. In the overwhelming majority of cases the particles of fresh snow are separate crystals, which grow with time and with the increase of the depth of the deposit.

The height of the capillary uplift and maximum water-retention capacity of freshly accumulated snow are much greater than that of old snow. This is connected on the one hand with the considerable absolute porosity and, on the other, with very small pores. The coefficient of airtightness is, as a rule, greater in the horizontal than in the vertical direction, which shows the weak development of the vertical migration of water vapour.

The layers of freshly deposited snow may be subdivided into three groups: deposited during blizzards, during low storms and during ground wind. The properties of each of these groups are given in Table 1.

TABLE 1

Properties of Freshly Deposited Snow

Indicator	Layer deposited during blizzard	Layer deposited during low storm	Layer deposited during ground wind
Density, gr/cm^3	0.36-0.42	0.44-0.50	0.38-0.44
Size of crystals, mm	0.28-0.40	0.15-0.25	0.25-0.35
Deviation from mean size of crystals, %	250-350	300-500	175-250
Hardness, kg/cm^2	1-3	6-10	3-6

Crusts are a characteristic feature arising in fresh snow. Four types have been distinguished: wind, rime, radiational and secondary layers of infiltration ice. The latter are due to the melting of snow at temperatures close to 0°C. Wind crusts form in winter (from May to August) under the influence of strong moist winds, during which a mechanical compacting of the snow and the sublimation of water vapour on the surface of the snow take place. Milk-opaque firn crusts about 1 mm thick and consisting of 0.25-0.30 mm chaotically oriented crystals arise.

Rime crusts are frequently observed on the surface of the snow sheet in autumn and winter. The crusts form during deep cyclones, when tiny drops of cooled cloud

water lying in this case on the surface of the snow sheet cover it with a thin film of ice. Inasmuch as during a cyclone the cloud line is at 500-600 metres, the maximum development of the rime crusts proceeds on slopes located at a distance of 15-35 kilometres from the coast. Rime crusts up to 3 mm thick consist of large (0.85 mm) and small (0.45 mm) crystals arranged in groups. The orientation of the crystals depends on the direction of the warm currents prevailing during the formation of the crust, but evidently the direction of these currents is determined by the configuration of the microrelief of the snow sheet and therefore does not follow any rule.

In the period from the second half of August until May crusts from under the joint action of the wind and radiational melting with a greater or lesser inclusion of ice in the crust, depending on the intensity of radiation. As the sun rises higher above the horizon, the radiational-wind crusts become radiational, but along the coast the wind continues to play an important part in the formation of crusts even in the period when melting is most intensive. Radiational crusts occur much less frequently on the periphery of the continent than in the central regions and particularly on the islands near the continent. This is linked with violent storms, during which the snow transported across the surface disperses and absorbs much of the solar radiation.

In more favourable conditions the radiational crusts attain a thickness of 2-3 cm. They consist chiefly of elongated pieces of ice with small section of firn between them. Almost all the crystals of the crusts are oriented at an angle to the plane «ab» in a northerly direction (Fig. 5), this direction of the optical axes exactly coinciding with the elongation of the crystals.

Fig. 5 — Stereogram of the orientation of the optical axes of crystals in the radiational crust. Conventional symbols: 1.—0-1%; 2.—1-3%; 3.—3-5%; 4.—5-8%; 5. — 8-12%; 6. — above 12%.

This structure of the crystals shows that their origin is the result of radiational melting that takes place under the influence of two causes. The first of these is the uneven relief of the surface of the crust consisting of bulging sections that may play the role of lenses collecting sunrays and concentrating energy at separate sections where the development of elongated ice is then observed. The second is that, as the investigations of M.A. Kuznetsov (1959) have shown, a large quantity of radiation penetrates through crystals whose optical axis is parallel with the rays of the sun; these crystals are best adapted for collecting energy and serve as the basis during melting.

Consequently, the following conditions may be used as a guide in defining radiational crusts: the optical axes of crystals, of which the radiational crusts consist, are oriented parallel to the main stream of radiation, i.e., northward, and together with the plane «ab» form an angle equal to the maximum height of the sun over the plane of the surface of the crust.

The snow sheet melts intensively only to an elevation of 400 metres above sea level, and there is some melting up to an elevation of 1,000 metres. In the central regions of the continent there is only a radiational melting of crystals. Four zones of ice-formation (Shumsky, 1955, 1957) are distinguished in Antarctica in conformity with the intensity of melting.

The strip directly along the coast, up to 1-1.7 kilometres wide, lies in the infiltration-congealation (ice) zone, where all the snow accumulated in the course of the winter turns into ice in summer. This zone is usually below the firn limit (Shumsky, 1955), but in Eastern Antarctica the firn limit is below sea level, which is proved by the presence of ice shelfs. That is why the presence of an ice zone is not a characteristic feature of the Antarctic. It forms only where by reason of the relief and as a consequence of strong winds the thickness of the layer of newly deposited snow is 100-200mm thick at the onset of the melting period and where the volume of melt water exceeds the volume of the pores of the unmelted snow.

A cold infiltration (firn) zone, in which the snow accumulated in the course of the winter turns in summer into infiltrative firn alternating with lenses of ice, stretches along a strip 10-13 kilometres from the shore (400 metres above sea level). At a distance of up to 100 kilometres (1,000 metres above sea level) there is an infiltration-recrystallization zone, where melting affects only an insignificant part of the summer layers of snow and does not affect the winter layers at all. Lastly, all the vast inland areas are in the recrystallization zone, where there is no melting with the exception of a radiational melting of crystals that leads to the formation of radiational crusts.

The determination of the annual volume of accumulation with the aid of pits sets an important task, which is to obtain precise criteria for delimiting the seasonal layers of the snow-firn. This is most difficult to do for the cold infiltration zone, because all the snow accumulated during the winter is modified as a result of melting, while where there are favourable conditions the water penetrates into the previous year's layers as well. However, a detailed examination brings to light the laws governing the structure of the snow in this zone too. The summer layers are well-formed lenses of ice. Where melting proceeds in an uninterrupted stream there is infiltration firn, whose crystals exceed 1 mm; beneath the infiltration firn, as a result of uneven melting of this part of the thickness, there is side by side with firn sections practically unchanged snow with relatively small crystals. The ice in such snow consists chiefly of tumour-like and, more rarely, of vertical ice inclusions.

The intensity of melting drops with height and the main summer level is the sedimentary layer that consists of 20-30 very thin (1-1.5 mm) alternating horizontal strata of ice and firn. The sedimentary layer attains a thickness of 12-15 cm. It forms as a result of radiational melting, during which water penetrates through the snow where the temperature is negative. Percolating through layers of various porosity and heat

volume, it saturates them to different degrees. Inasmuch as the primary wind layer is most often oriented horizontally, the sedimentary layers lie horizontally. Various stages of a sedimentary layer, forming a single genetical series from stratified firn to lenses of pure stratified snow without inclusions of firn occur in the snow firn sheet depending on the intensity of the melting.

M.A. Kuznetsov (1960) believes that the sedimentary layer takes shape as a result of a combination of melting and the drifting of snow. The snow carried during the night prevents the infiltration of water and each stratum of ice is the result of melting in the course of one day and is always separated from the stratum above it by at least a this layer of snow covering the surface during the night. The strata of ice are arranged one above the other at intervals of 2-3 mm to 3-4 cm, which corresponds with the accumulation of snow for several days, in the course of which there has been no melting because of the drift of large masses of snow or because of cloudy weather. The reason for the arrangement of the strata one above the other is also that a decrease of the albedo of the surface of the snow in the area of primary melting favours melting on the next day. The moisture that forms under such conditions does not penetrate down and, consequently, the sedimentary layer grows upwards.

Exact indicators determining the winter and summer layers of snow have been found for the infiltration-recrystallization and recrystallization zones of the formation of ice. The winter layers are, as a rule, harder, have a greater density and, consequently, a smaller porosity, small crystals, a high level of capillary uplift and wind crusts. The features of the summer layers are inconsiderable hardness and density, great porosity, relatively great crystals and, in connection with this, a low level of capillary uplift, radiational crusts and strata of ice formed during melting.

Melting proceeds irregularly in the course of the summer and the physical properties of the summer layers are more varied than those of the winter layers. As a rule, the appearance of water in a liquid phase leads to its migration into the lower levels. Here we observe the formation of washing away layers, whose density decreases, and in-washing layers, whose density increases: 0.368 and 0.442 or 0.442 and 0.467 gr/cm^3 respectively.

In the summer layers the redistribution of matter through the migration of water vapour leads to the formation of loose levels with an inconsiderable density—0.32-0.35 gr/cm^3. The floating snow in the loose levels has a well-defined vertical pillar-like texture. Here the feature is the elongated point of juncture between crystals arranged in a vertical direction. However, measurements showed that this elongation is not connected with the structural peculiarities of the snow. The crystals are isometric: their length horizontally is 0.88 mm, and vertically 0.93 mm. The optical axes of the crystals in any of the layers lying at a small depth are oriented chaotically.

Mainly two processes are observed in the upper part of the snow-firn layer: sinking, that leads to the snow becoming compact, and sublimatory re-crystallization, as a result of which the crystals grow larger. These processes are varied in the different zones of the formation of ice, as Table 2 shows.

TABLE 2

Compacting of Snow and Enlargement of Crystals in Different Zones of the Formation of Ice

	cold infiltration zone, 3 km. from the coast, 120 m above sea level		infiltration recrystallization zone, 43 km from the coast, 780 m above sea level		recrystallization zone, Komsomolskaya Base, 870 km from the coast, 3,540 m above sea level	
Year	density gr/cm³	size of crystals, mm	density gr/cm³	size of crystals, mm	density gr/cm³	size of crystals, mm
1957	0.420	0.29	0.415	0.25	0.414	0.25
1956	0.543	1.28	0.436	0.69	0.345	0.41
1955	0.560	0.72	0.470	0.82	0.372	0.47

The different intensity of the accumulation of snow and of the processes of development of the snow-firn sheet causes the firn to turn into ice at various depths. Seismic measuremets taken by the Soviet Antarctic Expedition showed that with distance inland there is a law-governed growth of the snow-firn sheet, which at Pionerskaya is 120 metres, at Komsomolskaya 137 metres and at the Pole of Relative Inaccessibility 164 metres.

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THE INTENSITY OF NOURISHMENT OF THE ANTARCTIC ICE SHEET

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SUMMARY

The nourishment of the antarctic ice sheet proceeds irregularly. On the coastal stretch it takes place through precipitation during the cyclons; in the central areas the snow has a form of hoar-frost. In the zone of the catabatic winds the drift during blizzards has great importance in the redistribution of the snow. For the most part, therefore, the discrepancy between the amount of snowfall and the deposit of snow is noted. It is established that due to the drift of snow during snowstorms the transference of snow masses takes place from the souther regions to the north but not farther than 50-100 km.

In the central areas of the mainland the annual amount of the nourishment averages some 50 mm of water. From the centre to the periphery the gradual increase of the nourishment occurs up to about 200 mm a year at a distance 350-400 km from the coast. Within 225-280 km there is the belt of maximum development of the catabatic winds, which carry away the snow from there to the coast; as a result the accumulation here does not exceed 30-50 mm a year. In the coastal regions the accumulation greatly differs from place to place depending on the direction and velocity of winds. As a rule it forms here 300-500 mm and in places increases up to 700 mm a year.

The intensity of snow accumulation on ice shelfs where the effect of catabatic winds is absent depends directly on the geographycal situation of the glaciers. The farther to the south is a glacier situated the smaller is the accumulation. The present active stage of the development of ice shelfs proves that the snow line in Antarctica lies below sea level even at the latitude of the polar circle. The existence of the parts of the ablation zone encountered up to the altitude of 1000 m are entirely due to strong winds.

Carrying away of snow from the mainland during blizzards is not great one (1,4 mln tons a year through one kilometer of coastline in the Mirny region). The ice westage proceeds mainly through calving of icebergs.

On account of the small amount of accumulation in the central areas of the mainland and the plane surface of the ice dome the movement of ice here is very slow. The age of the ice forming the central areas, therefore, is very great (the beginning of the glacier period) whereas along the external edge the ice sheet is formed mainly by considerably younger ice. By way of calving of icebergs the ice formed in the lower part of the slope of the ice sheet is spent. There is every reason to believe that the ice from the centre of mainland towards the coast comes in quite small amount.

RÉSUMÉ

L'alimentation de la couverture glaciaire de l'Antarctide procède irrégulièrement. Sur la ligne littorale elle s'effectue par les précipitations, dans les parties centrales du continent — c'est le givre, qui représente les précipitations principales. Dans la zone des vents de déversement une grande importance dans la redistribution de la neige joue la translation par tempêtes de neige. Voici pourquoi dans la majorité des cas, l'on observe une disparité entre la quantité de la neige précipitée et de celle qui est déposée. L'on a établi que grâce à la translation de la neige par tempête, il y a lieu un déplacement de masses de neige des régions plus méridionales vers le nord, mais pas plus loin que 100-150 km.

Dans les régions centrales du continent, la valeur annuelle de l'alimentation égale près de 50 mm. Du centre à la périphérie a lieu une hausse graduelle de l'alimentation jusqu'à 200 mm par an à une distance de 350-400 km du littoral. Dans les limites de 225-300 km se trouve la zone de développement maximum des vents de déversement qui apportent d'ici la neige vers le littoral; en résultat l'accumulation annuelle ici n'excède point 30-50 mm. Dans les régions littorales l'accumulation de la neige diffère grandement de place en place en fonction de la direction et de la vigueur des vents. De règle, elle égale ici 300-500 mm et s'accroît par place jusqu'à 700 mm.

L'intensité de l'accumulation de la neige sur les glaciers de shelf, où l'action des vents de déversement est absente, dépend directement de la situation géographique

des glaciers : tant plus au sud est le glacier, tant plus petite est l'accumulation. Le stade actif actuel du développement des glaciers de shelfs confirme que la ligne de neige dans l'Antarctide se trouve au niveau de la mer, même à la latitude du cercle polaire. La présence de parcelles de la zone d'ablation, qui se rencontrent jusqu'à une altitude de 1000 m, est définie entièrement par des grands vents.

La décharge de la neige du continent, durant les tempêtes de neige, n'est point grande (1,4 mln *t* par an à une distance d'un km de la ligne littorale dans la région de Mirny); voici pourquoi les données de Levé sont excessivement exagérées. La consommation de la glace se réalise, principalement, grâce au détachement des icebergs.

Dû aux petites valeurs d'accumulation dans les régions centrales du continent et à la superficie plane de la surface du dome glaciaire, le mouvement de la glace ici est insignifiante. C'est pourquoi la glace qui forme les régions centrales, est très âgée (début de la période de glaciation), tandis qu'à la bordure extérieure le continent est formé par de la glace beaucoup plus jeune (selon les données de P. A. Schoumsky. Par détachement des icebergs est consommée la glace, formée sur la partie inférieure de la pente continentale. Il y a toute raison à supposer que la glace du centre du continent atteint le littoral en quantités relativement petites.

In calculating the intensity of nourishment of the ice sheet we deal first and foremost with the mass balance of the surface, i.e., with the summary result of the action of all the processes taking part in the formation of the layer of firn. The debit of this balance is composed of precipitation, and the formation of increasing deposits and accumulations of snow carried from other areas as a consequence of storms. Expenditure of matter on the surface takes place as a result of the transportation of masses of snow during storms, evaporation and run-off.

The main difficulty in determining the intensity of nourishment of the Antarctic ice sheet is that the existing methods make it almost impossible to distinguish new-fallen snow from drifting snow. As the investigations at Mirny have shown, the quantity of snow measured by precipitation gauges does not in any way correspond with the quantity of deposited snow, as this is connected with the constant strong winds. But no regular connection between deposits of snow and the measurements shown by the precipitation gauge ([30]) were established even in the relatively peaceful conditions on the Maudheim ice shelf. The absence of reliable methods of determining the annual accumulation has frequently led to totally incorrect evaluations and, in connection with this, to erroneous conclusions about the tendency of the modern glaciation of the Antarctic to develop ([16]).

Reliable methods of determining the seasonal layers of snow for the different regions of the Antarctic have now been developed ([17]). With the help of data from pits this enables us to determine the magnitude of the annual net accumulation of snow and thereby to avoid determining the components of the mass balance of the surface, which is difficult to do, and, as a rule leads to very inaccurate results. However, in order to elucidate the laws governing nourishment in the different regions of the continent it is necessary to give a brief evaluation of these components.

1. Precipitation

The feature of the atmospheric circulation in the Antarctic and along its periphery is that masses of warm and moist air constantly move towards the continent from the north. As the air moves southward, the moisture in it condenses and falls in the forms of snow, the greatest snow-falls being at an elevation of 400-1,000 metres above sea level. Along the coastal strip snow falls chiefly during cyclones and is accompanied by easterly and north-easterly winds. Thus, the main mass of snow in this area is accounted for by cyclonic precipitation.

Moisture-containing masses of air also move deeper into the continent and cause

considerable snowfalls up to an elevation of at least 3,000 metres. This precipitation loses its intensity as it moves farther inland, although some deep cyclones penetrate considerably farther; they have been registered at all inland stations. Consequently, there may be cyclonic precipitation within the continent as well. Anticyclonic precipitation are of great importance in the central regions of the Antarctic; they are produced not by passing cyclones but by currents of relatively warm air moving constantly towards the surface. Due to the strong surface inversion during this process, a sublimation of water vapour takes place and this leads to the formation of small crystals of ice in the clear sky, which fall slowly to the surface and form the main mass of snow [3].

2. Sublimation and Evaporation

Moisture sublimation on the surface and hoar-frost forms as a consequence of the considerable difference between the temperature of the cold surface of the ice and the air over it. This process is very intensive on ice shelfs, and in the central regions of the continent. At Vostok Base hoar-frost forms almost all year round. Only in the winter (from June to September) is its formation hindered by winds blowing at a velocity exceeding 4 metres per second. As a rule, hoar-frost does not form along the sea coast because of the strong catabatic winds.

Evaporation of snow is especially intense in the regions where catabatic winds develop and sometimes acquire the character of real föhns. The presence of zones of ablation on the continent is entirely determined by winds blowing away the snow and causing increased evaporation. In extreme conditions, in Terre Adelie, the ablation zone extends for 11 kilometres from the coast to an elevation of 500 meters above sea level, and there, as a consequence of evaporation and the wearing away action of the drifting snow, the ice level drops by an average of 15-18 cm a year, which is equivalent to 100-120 mm of water [19]. Other places do not have a solid zone of ablation but there are separate sectors (up to an elevation of 1,000 metres) that conform with the relief and with the direction of the dominant winds.

N. P. Rusin [8] has calculated that evaporation totalling 220-250 mm goes on all year round in the coastal regions exposed to catabatic winds. In the interior, on the contrary, a sublimation of moisture is observed on the surface. However, because of the low temperatures and the small content of water vapour in the air, the total magnitude of precipitation, thanks to sublimation, is usually 15-20 mm and does not exceed 40 mm a year.

3. Drifting snow

To date, the problem of drifting snow is the least clear. Mainly three questions are debated: the ways snow drifts in the continent, the distance it drifts and the transportation of masses of snow from the continent to the sea. The reason these questions have not been settled is that we do not have a good method or the measuring instruments.

In accordance with the main directions of the winds in the continent, snow drifts in two directions: from east to west and from south-east to north-west. In the first case the stream of snow and wind contains chiefly snow falling at the given moment and by virtue of the laws of drifting snow it cannot be carried for any great distance. Moreover, during easterly winds snow moves parallel with the coast of the continent. Because of this the easterly drift is not of great importance in the redistribution of snow in the continent.

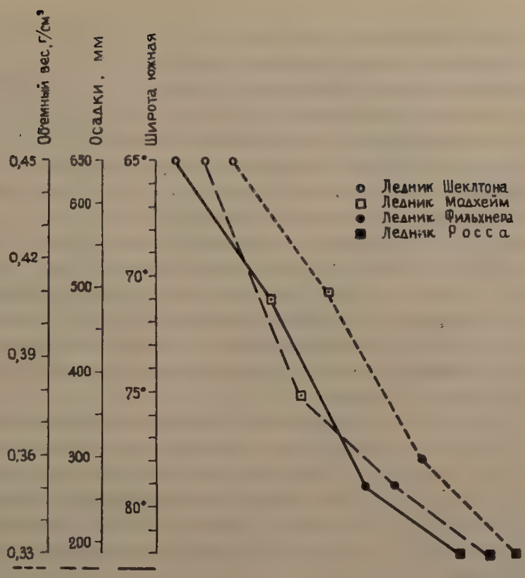


Fig. 1 — Dependence of the intensity of nourishment and the density of the snow upon the ice shelfs on the geographical position.

Catabatic winds that cause considerable masses of snow to move down the slope of the ice sheet are predominant in a strip up to 600 kilometres wide along the coast. On the basis of the laws governing the height of snowdrifts, the density and hardness of the surface layer of the snow sheet, P.A. Shumsky advanced the conjecture that there is an axis of the belt of catabatic winds at a distance of 250-280 kilometres from the coast (9). Although here the velocity of the catabatic winds is somewhat less than in the lower part of the slope of the ice sheet, they blow continuously throughout the year, while along the coast their strength abates considerably during the warm months. That is why the general influence exerted by the winds on the snow sheet is the greatest in a strip 250-280 kilometres wide along the coast. The winds blow the new snow off the surface, carrying it to the more northerly regions, and form vast areas where the snow that had accumulated earlier is dispersed. Here P.A. Shumsky found firn several decades old on the surface. The crystals were properly oriented along the vertical direction.

The character and distance of the transportation of snow by catabatic winds depends on the configuration of the relief and on the time of the year. On the basis of an analysis of the form, size and crystallographic properties of the transported granules, it has been established that the snow is carried over as distance up to 50-100 kilometres. The following three facts show that snow is carried from the zone of catabatic winds to the coastal strip.

The precipitation of mainly crystals of a lamellar type of growth are a characteristic feature of the coastal strip of the continent. However, pillar-like crystals begin to be predominant in the snow sheet towards mid-winter. This could be explained easily by the general fall of temperature and the rare occurrence of deep cyclones. However during observations of the precipitation it was noted that there were considerably fewer pillar-like crystals than in the pits. This means that the sharp increase

of the quantity of pillar-like crystals in the snow in mid-winter is linked up with the transportation of particles of snow from the hinterland.

Within 20-40 km from coast it was noted that there was a larger quantity of snow in the snow pits in the summer layers than in the winter layers. The explanation for this is that as a result of the high temperatures predominating in this area in the summer, the snow carried from the interior of the continent piles up and cannot be transported farther towards the coast and is thrown into the sea. In winter the situation changes and the snow is carried farther towards the shore and to the sea.

The accumulation of snow directly on the coast and 6 kilometres away from it is analogous throughout the year with the exception of the winter months. Towards the end of winter no increase in the snow sheet was observed along the coast, but 6 kilometres inland the height of the sheet increased by 33 cm. This accumulation of snow at 5-10 km from the coast, when there was an almost total absence of cyclonic activity, shows that in winter snow is transported continuously and is deposited closer to the sea shore than in summer.

Huge masses of snow are carried from the continent to the sea during low snowstorms. However, Loewe's ([18]) estimate that 18 million tons of snow are carried across each kilometre of the coastline is much too high or is characteristic only of such extreme conditions as occur on Terre Adelie. Investigations in the region of Mirny show that about 1,500,000 tons of snow is carried a year ([4]), and this enables us to assert that though drifting snow takes part in the mass balance of the glacier it is not of essential importance.

4. Melting

Through the investigations at Mirny, Terre Adelie ([19]), McRobertson Land ([23]) and Queen Maud Land ([28]) it has been established that melting is intensive at an altitude of up to 400 metres above sea level and is insignificant at an elevation of up to 1,000 metres. But in all cases there is no drainage of melt water, as it completely infiltrates into the lower layers of snow. Drainage of melted snow is observed only at the barrier, but it has no effect at all on the mass balance of the surface.

In the continent, the above factors manifest themselves to varying degrees and lead to a sharp differentiation of the nourishment of the ice sheet. We may assume that there are two laws governing the changes in the nourishment and they are linked up with the general circulation of the atmosphere in this part of the globe: latitudinal and meridional. 1. The intensity of nourishment decreases with distance inland as a result of the gradual abatement of cyclonic activity. 2. Because of the meridional high pressure crests and the regions of low pressure depressions between them where slow-moving cyclones reign ([6]), as well as the sections of the slope along which cyclones usually move to the continent, the intensity of nourishment turns out to be different along one and the same latitude, but with distance in land this difference is smoothed out. Both laws are broken by the influence of catabatic winds.

Available data enable us to give an estimation of nourishment for the whole of the Antarctic.

Ice shelfs are situated in regions with the greatest precipitation. They are formed by local precipitation brought by the frequent cyclonic invasions. No catabatic winds blow over them. On the basis of measurements taken with stakes and in pits ([31, 32]), the annual accumulation of snow on the Ross ice shelf was found to be 180 mm. The data obtained along the traverse of the Filchner glacier in 1957-58 ([11]), give the accumulation as 270 mm a year. On Maudheim ice shelf measurements taken in a deep pit ([27]) and the study of an uncovered iceberg ([28]) showed the accumulation

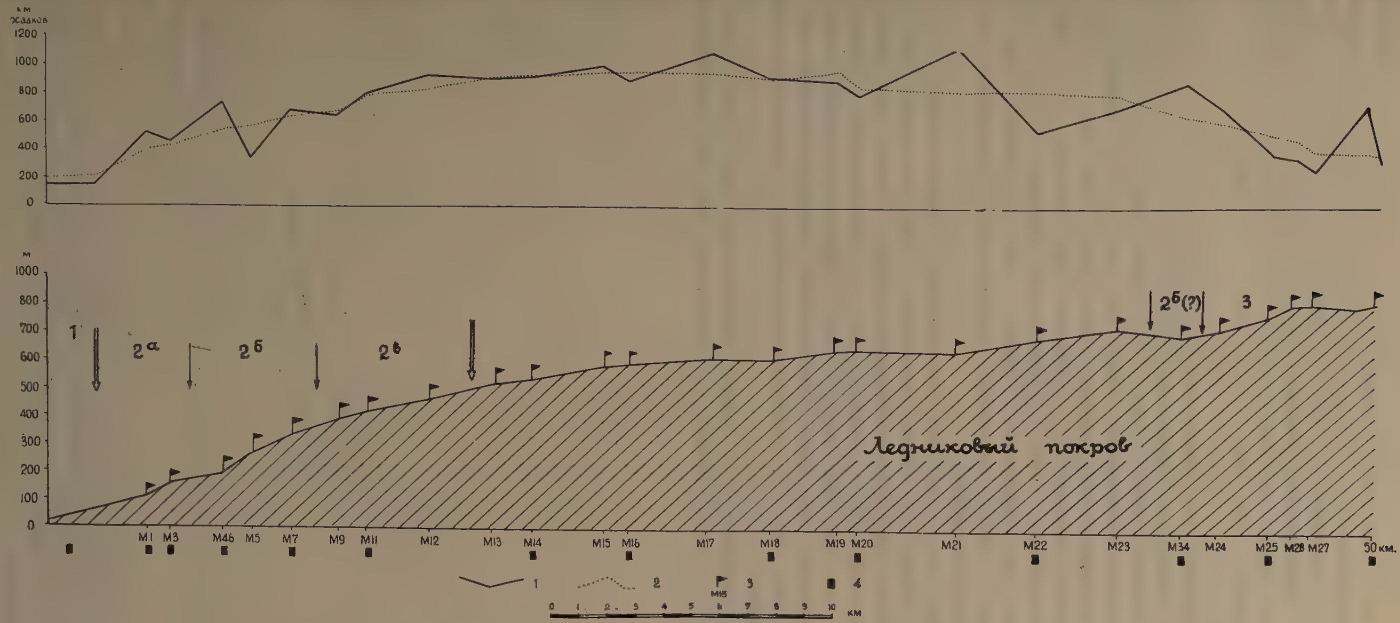


Fig. 2 — Volume of accumulation in the lower part of the slope of the ice sheet in the region of Mirny. Conventional symbols: 1.—volume of accumulation in 1957; 2.—accumulation curve, obtained on the basis of the mean for five points; 3.—location of stakes on the slope; snow pits.
The zones of formation of ice: 1. infiltration-congelation; 2. cold infiltration: *a*) lower part, *b*) middle part, *c*) upper part; 3. infiltration-recrystallization.

to be 370 mm a year. Lastly, our investigations on the Skackleton ice shelf gave the accumulation as being about 700 mm. As Fig. 1 shows, the intensity of nourishment of ice shelfs by precipitation is directly dependent on the latitudinal location. The farther south a glacier is, the less precipitation it receives. The reason for this is the increased distance from the main paths of the movement of cyclones. An analogous dependence is observed in the case of the density, which decreases in the more southerly glaciers.

The accumulation of snow differs greatly from place to place in the lower part of the slope of the ice sheet and depends on the direction and velocity of the winds. As a rule, it is 300-500 mm and increases to 700 mm here and there. The profile of accumulation in the 50-kilometre coastal strip in the region of Mirny is typical in form although it has extraordinarily high absolute values (Fig. 2). Maximum accumulation of snow is observed at a distance of 15-35 kilometres from the coast at an elevation of 400-550 metres above sea level, which corresponds with the usual height of the clouds. Higher up the slope the accumulation decreases: at the 50th kilometre there is an annual accumulation of 310 mm, at the 200th kilometre only 120 mm and at the 250-280th kilometre less than 100 mm.

Strong catabatic winds distort the correct picture of the accumulation of snow. On McRobertson Land the result of these winds is that in an area from 26 to 80 kilometres from the coast the accumulation amounts to 50 mm, in the area from 80 to 192 kilometres 100 mm, in the area from 192 to 250 kilometres 170 mm, and in the area from 250 to 320 kilometres 200 mm ([23]). Because of catabatic winds the accumulation of snow on Terre Adelie does not exceed 200 mm beginning from the 11th kilometre from the coast ([19]). For the same reason only 131 mm of snow accumulates at Wilkes Base ([12]).

Decreased accumulation due to a smaller content of moisture in the air and the distance from the sea coast is observed at an elevation from 1,000 to 2,000 metres

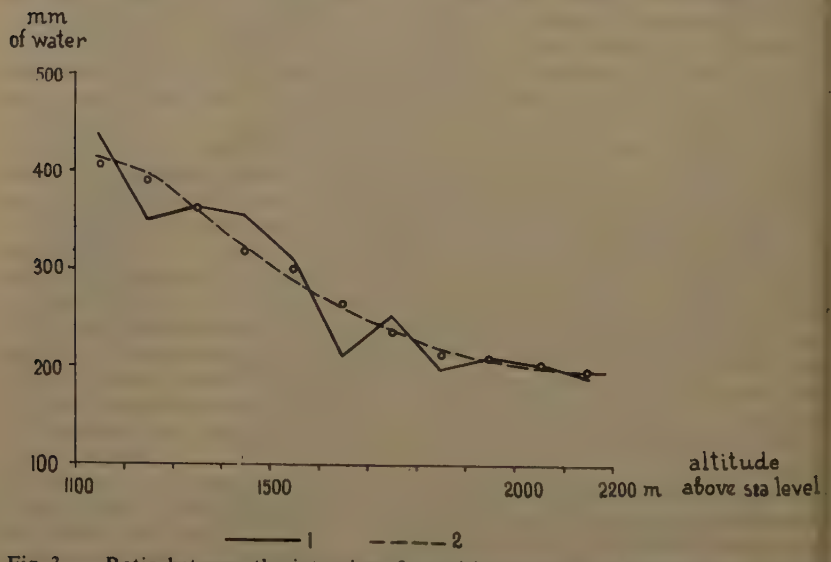
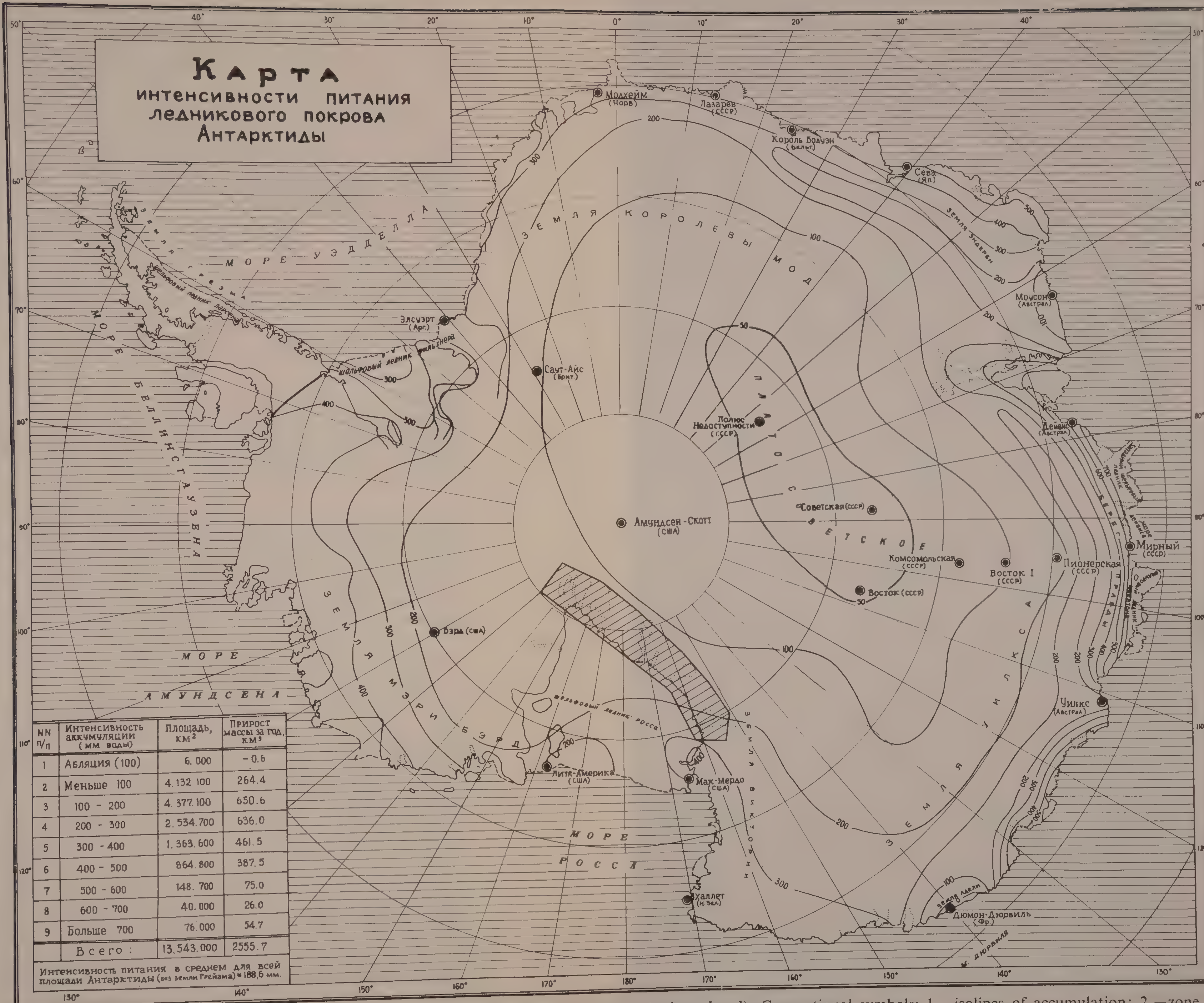


Fig. 3 — Ratio between the intensity of nourishment and the absolute elevation of the surface of Mary Byrd Land. Drawn by the author in conformity with materials (10). Conventional symbols: 1.—mean volume of accumulation for elevation intervals of 100 metres; 2.—accumulation curve obtained on the basis of the mean for five points.



NN п/п	Интенсивность аккумуляции (мм воды)	Площадь, км²	Прирост массы за год, км³
1	Абляция (100)	6. 000	− 0.6
2	Меньше 100	4. 132 100	264.4
3	100 - 200	4. 377. 100	650.6
4	200 - 300	2. 534. 700	636.0
5	300 - 400	1. 363. 600	461.5
6	400 - 500	864. 800	387.5
7	500 - 600	148. 700	75.0
8	600 - 700	40. 000	26.0
9	Больше 700	76. 000	54.7
	Всего:	13. 543. 000	2555.7

Интенсивность питания в среднем для всей площади Антарктиды (без земли Грейама) = 188,6 мм.

Fig. 4 – Chart of the intensity of nourishment of the Antarctic ice sheet (excluding Graham Land). Conventional symbols: 1.—isolines of accumulation; 2.—zone of the greatest influence of catabatic winds; 3.—region of highland glaciation with complex accumulation.

above sea level when strong catabatic winds do not blow. This phenomenon is clear-cut on the graph for Mary Byrd Land (Fig. 3).

Accumulation fluctuates considerably when the even surface of the ice sheet is broken by the emergence of basic rock. Such, for instance, is the region of the mountain range stretching along the Ross ice shelf.

Southwards, the surface levels out and the conditions for the accumulation of snow become stable. Above an elevation of 3,000 metres, the meridional differences in the nourishment of the various regions of the Antarctica smooth out and an identical decrease in accumulation takes place throughout the continent at its highest points. The periphery of the central region of the Antarctica receives about 100 mm of precipitation a year: Vostok-1 — 100 mm, Komsomolskaya — 80 mm ([3]); South Ice — 100 mm route of the Fuchs crossing — 80-90 mm ([21]); Amundsen-Scott — 60-90 mm.

The highest sections of the Antarctica ice sheet (Plateau Sovetskoye) have a mean annual accumulation of about 50 mm. Stake observations in 1958 showed that about 30 mm of snow accumulates at Vostok and Sovetskaya. A comparison with the data obtained at other stations gives grounds for considering these magnitudes as being at least 1.5 times less than the average.

On the basis of available data we have drawn up a chart showing the intensity of nourishment of the Antarctica ice sheet (with the exception of Graham Land). The data has been taken from Soviet, American, the British Trans-Antarctic, the Norwegian-British-Swedish and the French expeditions (Fig. 4). The appended chart does not claim to be very exact because there is still a shortage of initial data. However, the available data is sufficient to reveal the actual magnitudes and the general geographic laws governing the nourishment of the ice sheet. In addition, they enable us to calculate the debit of the mass balance of the ice sheet by the planimetric method.

Most of the recent analogous calculations were made on the basis of an analysis of single meridional profiles, a factor which introduced two important errors into the final result. 1. The meridional profiles reflect the picture of the latitudinal changes of accumulations on the territory of the continent with sufficient accuracy, but take no account at all of local differences, which are well outlined on the appended approximate chart. 2. The meridional profile is a linear value, while in calculating the mean magnitude of nourishment it is necessary to operate with areas with different quantities of accumulation.

The calculation made by the planimetric method yielded a mean value of nourishment for the whole of the Antarctica: $\sim$ 190 mm of precipitation. This value is bigger than the earlier estimates: 70 mm ([22]); 90 mm ([15]); 100 mm ([20]); 130 mm ([21]); 140 mm ([24]) and only less than that of Kosack- 201 mm ([16]).

Data on accumulation are available for different parts of the continent to a varying extent. The situation is worst of all with regard to Wilkes Land and Enderby Land, for which there is inadequate data. In the latter regions, the isolines of equal accumulation have been drawn on the basis of the general laws governing the accumulation of snow and by analogy with adjoining regions. If for doubtful regions the magnitude of accumulation is cut by 100-200 mm, then in the computation for the territory of the entire continent this will yield 180 mm a year. The total evaluation will thus change insignificantly.

The importance of this mean value is conditioned by the intensive accumulation on the periphery of the continent. A certain disproportion, compensated by movement is caused by the uneven nourishment of the outlying and interior regions of the continent. A small accumulation in the central regions has resulted in a very flat surface, gentle slopes, and, as a consequence of this, an extremely slow movement ([9]). That is why the ice in the central regions of the continent is very old (early Ice Age), while along the edge of the continent the ice is considerably younger.

As a result of the small values of nourishment and the slowness of movement is the central regions of the continent, very small quantities of ice are borne from the centre to the coast. The ice on the ice shelfs and in the lower part of the slope of the ice sheet is expended chiefly in the shape of icebergs.

Having a trustworthy value for the income of the mass balance of the ice sheet, we shall try to evaluate the expenditure, which consists, on the one hand, of the movement of the ice sheet and the breaking away of icebergs, and, on the other, of melting in the lower surface of the shelf glaciers. Regrettably, we only have fragmentary data about the velocity of the movement of the Antarctic ice sheet and other of processes taking place on the edge of the ice—water; the work done in the period of the International Geophysical Year added little to this.

The following table may be drawn up on the basis of data available in literature on the rate of movement and thickness of the ice along the edge of the continent:

Type of glaciers	length of coastline, km	rate of movement m/year	thickness at the edge m	density of ice gr/cm^3	expenditure of matter, km^3
Undifferentiated edge of the ice sheet	12 100	100	250	0,88	270
Ice shelfs	8 600	360	200	0,84	520
Ice streams	1 800	650	400	0,90	420

On the basis of recent investigations it has been proved that most of the ice shelfs melt although it is not impossible that ice accumulated in separate places where there is no circulation of water. However, such places are very few. On Maudheim glacier 7 cm of ice, which is equivalent to 62 mm of water ([29]), melted in the course of six months. If we take about 100 mm as the mean annual melt, then with the total area of the ice shelfs being 1,2 million square kilometres, this will be equivalent to 120 cubic kilometres.

The Antarctica (without Graham Land) thus expends 1 330 cubic kilometres of water a year, while the debit amounts to 2 550 cubic kilometres. It may therefore be asserted that at present the Antarctic ice sheet is increasing. The same conclusion has been drawn by most of the present-day investigators of the Antarctica: Loewe ([20]), Mellor ([24]), Lister ([21]).

Two important conclusions may be drawn from the obtained data. It is evident that of the two components of the balance, the magnitude of expenditure is not trustworthy. It may be surmised that the rate of movement of the ice sheet is very fast in separate places and that thus the Antarctic is discharging excessive ice. A study of aerial photographs testifies to the considerable rate of movement of the ice ([14]). However, a simple computation shows that if the mean thickness of the ice at the edge of the continent is taken as 250 metres and its density as 0,89 gr/cm^3, then a rate of movement of 500 metres a year is necessary at the edge of the entire continent to maintain equilibrium.

It remains to surmise that at the present time the ice sheet is increasing in volume. At the same time indisputable facts show that in the coastal belt of the continent the

ice retreated in the course of the 19th century and early 20th century. This is expressed in the lowering of the surface and in the retreat of the ice barrier.

These seemingly contradictory results are explained by the character of the changes in the present-day climatic conditions in the world, the chief feature of which is the general rise in temperature, which has been measured with instruments in the Antarctic as well. The rise in temperature is activizing atmospheric circulation and intensifying the meridional exchange of masses of air. Warmer air with a greater content of moisture flows over the continent. However, the rise in temperature by a few degrees does not cause any ablation in the interior of the continent, while the increased moisture is leading to more abundant snowfalls.

At the same time, the warmer climate in the coastal regions, where in summer the temperature is around 0°C, is bringing about a considerable intensification of ablation and is forcing the ice barrier to retreat, a process that has been observed in the past century.

The increase in the mass of the Antarctic ice sheet is leading to a higher rate of movement of the glacier and after a span of time this will find a reflection in the periphery. Reference to the present-day stationary position of the external section of the ice sheet is appearing in literature ([1, 25, 26]).

The most dynamic Antarctic glacial formations—ice shelfs—likewise testify to the increase in nourishment and, in a number of cases, to the fact that this supply is greater than the expenditure. A feature of the present time is the formation of ice shelfs from old coast ice ([5, 7]). Today the barrier of the Shackleton and Western ice shelfs are several tens of kilometres farther north than it is shown on American maps compiled on the basis of material obtained up until 1947. In places the ice shelfs tend to become part of the continental ice sheet, e.g. the core of Shackleton glacier to the south of Masson Island ([2]). The ice shelfs are at present thus in an active stage of existence and in the climatic conditions obtaining at present do not show any tendency to decrease. On the contrary, in a number of cases they are increasing both in area and in thickness.

Thus, on the basis of a study of the intensity of nourishment of the Antarctic ice sheet and by comparing the supply and expenditure of matter we may speak of the present increase of the volume of ice in Antarctica, but to assert this unconditionally it is necessary to study the movement of the ice sheet in the continent in detail and this is easiest with the aid of repeated aerial photographs.

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GROWTH AND THERMAL STRUCTURE OF THE DEEP ICE IN BYRD LAND, ANTARCTICA *

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SUMMAY

Instead of starting with an initially dry, below sea-level basin undergoing glaciation, as was done in an earlier model, the new computations start with an open channel connecting the Ross and Bellingshausen Seas. The period of glaciation begins as the sea in the channel freezes permanently and acquires accumulation, both from local precipitation and transport from the adjacent mountains.

With an average annual accumulation of 20 cm and freezing from below, the time required for the growing ice shelf at 80°S, 120°W to become grounded at a thickness of 309 meters is computed for three different models, yielding 7,100 years, 6,000 years and 4,300 years. Using the middle value, an analysis is made of how the excess mass and energy are dispersed by the lateral spread of the ice and breaking off in bergs at the edges. After grounding of the ice and cessation of phase changes at the bottom, the additional time required for ice to build up to its observed thickness of 2,300 meters is 23,100 years. The temperature at the ice bottom remains at the pressure melt point even after grounding of the ice, and geothermal heat flux at the deep-sea rate of 10^{-6} cal cm $^{-2}$ sec^{-1}, does not contribute materially to melting. The temperature in the ice declines sharply with height above the bottom, and becomes practically isothermal in the top 400 meters. The observed temperatures in the 310 meter deep hole at Byrd Station show a slight decrease with depth. If this is interpreted solely in terms of a climatic change, it would mean a warming of 1.56°C/1000 years. If it were interpreted solely as an effect of ice motion down-slope to warmer temperatures, then using down-slope speeds of 10,14, 20, 123.5 meters yr^{-1} and an observed vertical temperature gradient of -0.0289°C/100 meters in the lowest 150 meters of the hole, the vertical temperature gradients in the surface layer can be computed. From these and a reference temperature at 121.92 meters depth, the average annual snow surface temperatures can be found. The 3 lowest ice speeds give temperatures with a few 0.1°C of the observed annual air temperatures of -28°C but the highest speed of 123.5 meters yr^{-1} gives temperatures 1.3° to 1.6°C too warm.

Finally, the computations are applied to the 4,300 meters of ice in the deep basin 200 kilometers east of Byrd Station. Here a maximum of 90,000 years of combined accumulation and freezing is required for the ice shelf to become grounded at a thickness of 1,500 meters and an additional 22,650 years for the remaining 2,800 meters to accumulate. Again, starting from the pressure melt point at the ice bottom, the temperature decreases sharply with height and becomes essentially isothermal in the top 1,000 meters.

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TRANSIENT TEMPERATURE DISTRIBUTIONS IN ICE CAPS AND ICE SHELVES

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Summary

Finite difference equations are established for the linear conduction of heat in ice caps and ice shelves subject to net accumulation and surface warming (whether due to lowering of the surface or climatic change). Stability considerations show that the central difference form of the equations has no stable solution, while for non-centered differences the grid interval must equal or exceed the square root of $2K_z\Delta t$, where K_z is the thermal diffusivity at depth z and Δt the length of the time step used in the numerical integration.

Results of numerical integrations with the digital computer CSIRAC at the University of Melbourne are presented for the following cases:

a) Station Centrale, Greenland. Heuberger's temperatures are found to be compatible with an ice movement at the rate of 10 m/year and a secular rise in air temperature at the rate of 2°C/100 years during the past 50 years.

b) Byrd station, Antarctica. The calculation suggests for the lowest temperatures a level slightly lower than that found by Wexler's treatment, and melting at the bottom of the ice.

c) Maudheim ice shelf. The temperatures in the 100 m borehole suggest bottom melting and a residence time of the ice of approximately 60 years.

d) Ross Ice shelf. The temperatures in the borehole through the shelf at Little America seem explicable only by the assumption of at first rapid and then extremely rapid melting, producing a reduction in thickness of the shelf by 50 % in the last 80 km from the ice edge. This thickness reduction is introduced into the calculations by permitting the temperatures in the lowest layer of the original mass of ice to rise above the freezing point, an artifice the full scope of which requires further study.

Zusammenfassung

Die strenge Lösung der Wärmeleitungsgleichung für eine Eisdecke mit Oberflächenerwärmung und Zuwachs zeigt, dass ein stationärer Zustand unter solchen Bedingungen selten erreicht werden kann. Die entsprechende Differenzengleichung muss in einer nicht-zentrierten Form benutzt werden, um die Stabilität der Rechnungen zu sichern, und wird für die folgenden Fälle numerisch integriert:

a) Station Centrale, Groenland. Die beobachteten Temperaturen lassen sich mit einer Eisgeschwindigkeit von 10 m/Jahr und einem Temperaturanstieg von 1 bis 2°C in den letzten 60 bis 70 Jahren vereinbaren.

b) Byrd Station, Antarktik. Die Rechnungen deuten auf eine Eisgeschwindigkeit von 30 m/Jahr. Die gemessenen Temperaturen sind mit verschiedenen Anfangsprofilen vereinbar, diese führen jedoch zu recht verschiedenen Resultaten am Boden des Eises, wenn seine Schrumpfung in Betracht gezogen wird.

c) Maudheim Schelfeis. Die beobachteten Temperaturen scheinen mit Schmelzen am Boden und einem Aufenthalt von etwa 50 Jahren in dem Schelfeis vereinbar.

d) Ross Schelfeis. Die Temperaturen in dem Bohrloch durch das Schelf in Little America scheinen nur verständlich wenn man annimmt, dass mit Annäherung and die Eiskante das Schmelzen zunimmt und die Eisdicke um etwa 20% in den letzten 80 km reduziert wird.

Im Ganzen bestätigen die Ergebnisse die Bedeutsamkeit des von Robin entdeckten Zusammenhanges zwischen der Bewegung und Temperaturverteilung grosser Eisdecken.

1. Introduction

It has been suggested (Radok 1959) that the effects of accumulation and spreading of an ice sheet on its internal temperature distribution (Robin 1955) might be considered jointly by studying the conduction of heat in a coordinate system moving

quasihorizontally with the ice. This is legitimate because the horizontal temperature gradients, although of the same order of magnitudes as the vertical gradients, are in general much more uniform than the latter and do not contribute appreciably to the Laplacian of the temperature and hence the temperature change by conduction. The latter can therefore be described by the relation

$$K \frac{\partial^2 T}{\partial z^2} - v \frac{\partial T}{\partial z} - \frac{\partial T}{\partial t} = 0 \tag{1}$$

where v is the vertical velocity of the ice (equal to the net accumulation for $z = 0$) and K the thermal diffusivity; and the increase in surface temperature due to the decreasing height of the ice surface towards the border (Robin 1955) is embodied in the boundary condition for $z = 0$

$$T_0(t) = T_0(0) + \alpha t \tag{2}$$

For the semi-infinite solid, constant α and initial temperature gradient A, and v and K independent of depth, the solution for a very similar system of equations has been given by Benfield (1949) who was concerned with the cooling and erosion of rising rock masses. Thus his equations differ from (1) and (2) in the signs of the term containing the vertical velocity v and of α. A detailed discussion of the solution for the spreadig semi-infinite ice cap with net accumulation (Radok 1959) will be given elsewhere. While not directly applicable to ice of finite thickness this solution provides a first pointer to the limits of depth and time within which the assumption of steady state heat conduction is not permissible.

As an example fig. 1 indicates for two net accumulation rates where in a depth-time coordinate system the vertical temperature gradient reaches certain fractions R_1 of its asymptotic value (cf. Radok 1959) given by

$$\gamma = -\alpha/v \tag{3}$$

which is identical with Robin's (1955) well known kinematic formula. As labelled the curves are valid for initially isothermal conditions. When instead the constant initial gradient is a fraction p of the final value given by (3) the ratio R_1 in fig. 1 becomes

$$R_2 = p + R_1(1 - p) \tag{4}$$

so that merely a re-labelling of the curves is necessary. For $p = -0.1$, corresponding to an initial gradient of 2°/100 m, roughly in accordance with the geothermal heat flux, and a realistic final gradient of −0.2°/100 m the curves in Fig. 1 would correspond to 12, 45, 78, and 89% instead of the values shown. Evidently the initial gradient has little effect, for practical purposes.

Fig. 1 demonstrates that steady state conduction is reached more rapidly with increasing net accumulation. For the smaller accumulation rate, a lower limit for the bulk of Antarctica, periods of the order of 4×10^4 years are required to bring the temperature gradient in the uppermost 300 m to within 20% of its steady state value. In this time the ice will normally move through a distance of the order of 400 km and change its surface slope, quite apart from climatic changes. Thus even on the simple assumptions implied in fig. 1 steady state conduction cannot often be realized.

2. Numerical solutions

For ice caps of finite thickness and ice shelves the strict solutions of (1) if they exist would be very cumbersome even for the simplest of boundary conditions. We

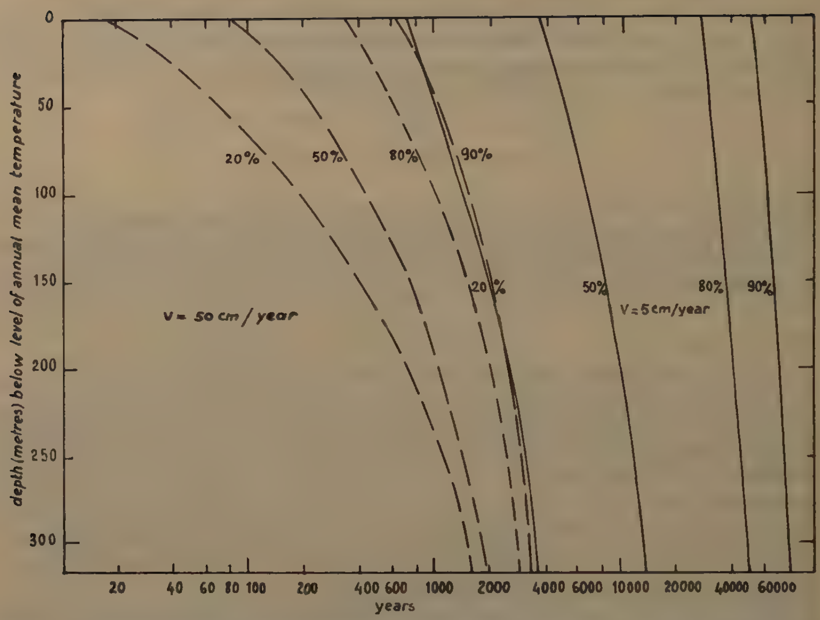

Fig. 1 — Isolines of R_1, the ratio of the transient and asymptotic vertical temperature gradients in a semi-infinite ice sheet subject to steady surface warming and net accumulation at the rate of v cm/year.

have therefore concentrated on numerical solutions. This made it possible to let the vertical velocity decrease linearly with depth, as suggested by theories of ice cap flow (Nye 1959) and to take into consideration also the change with density in the thermal diffusivity K, represeted by

$$K = k/c\varrho = 0.01415\ \varrho \qquad (5)$$

where $c = 0.48$ is the specific heat of ice at $-13°$ and k its conductivity, here assumed as following Abel's relation $k = 0.0068\ \varrho^2$.

Equation (1) then takes the form (1)

$$K_z \frac{\partial^2 T}{\partial z^2} + \left(2 \frac{\partial K_z}{\partial z} - v_z\right) \frac{\partial T}{\partial z} = \frac{\partial T}{\partial t} \qquad (6)$$

with the finite difference equivalent

$$K_z(T_{z+1,t} - 2T_{z,t} + T_{z-1,t})/\Delta z^2 +$$
$$+ [(K_{z+t} - K_{z-1})/\Delta z - V_z]\,(T_{z+1,t} - T_{z-1,t})/2\Delta Z = \Delta T/\Delta t \qquad (7)$$

For the right hand side, we have the alternative of using a central or non-central diffe-

(1) In the original version of the paper the factor 2 was inadvertently omitted from equation (6) and its finite difference equivalent (7). This had no appreciable effect on any of the transient temperature distributions but exaggerated the curvature of the steady state profiles in fig. 4, introducing errors of up to 0.5°C. The writers are grateful to Professor J.C. Jaeger for pointing out this error.

rence form, viz.

$$\Delta T/\Delta t = (T_{z,t+1} - T_{z,t-1})/2\Delta t \text{ or } (T_{z,t+1} - T_{z,t})/\Delta t \tag{8}$$

However, a study of the computational stability of (7) along the lines of Hildebrand (1952), sect. 3.23) showed that an essential requirement for stability is that the factor of $T_{z,t-1}$ must be positive or zero. Hence the central difference form (8) cannot be rendered stable by any choice of Δz and Δt. For the non-central form the factor is zero and the stability criterion turned out to be

$$\Delta t \leqslant \Delta z^2/2K_z \tag{9}$$

With $K_z \leqslant$ 40 m²/year this becomes approximately

$$\Delta z \geqslant 9\Delta t^{1/2} \tag{10}$$

Combinations satisfying (10) and used in the following are $\Delta z = 50$ m, $\Delta t = 20$ years (ice caps) and $\Delta z = 25$ m, $\Delta t = 5$ years (ice shelves).

As surface boundary condition we shall in general take a linear rise in surface temperature; occasionally the rate of warming will be varied so as to simulate secular air temperature (or surface inversion frequency) trends. At the bottom of the ice in most cases a constant heat flux from below is assumed, although again not necessarily at one and the same rate throughout an entire integration. Since with one exception the temperature observations now available for comparison with computed temperatures have been made near the surface of the ice they could not be expected to provide a discriminating check on the lower boundary condition.

The numerical integrations were programmed by one of us (Jenssen) for the electronic computer CSIRAC, a single-address binary digital computer with a command time of 220μ sec and a high-speed storage capacity of only 768 words of 20 digits each. This made it necessary to overwrite the initial part of the program by later commands; nevertheless sufficient flexibility was preserved to permit changes of the surface warming rate (though not of the grid size) during any calculation. The program was found to operate effectively at a speed of 12,000 $\Delta t/(N - 2)$ years of conduction per hour, where N is the number of grid points. Thus 1 000 years of conduction in a 3 000 m ice cap ($N = 60$) required fifteen minutes while 100 years of conduction in a 250 m ice shelf ($N = 10$) took a little over one minute.

The factual framework for the calculations was provided by measurements at two ice cap stations, Station Centrale in Greenland and Byrd Station in Antarctica, and at two ice shelf stations, Maudheim and Little America. The thermal diffusivities as function of density are shown in fig. 3 inset; they were computed from the densities reported for Byrd and Maudheim, using the density-depth relation derived by Robin and Schytt (cf. Schytt 1958)

$$0.917/(0.917 - \varrho)\, d/dz(1 - \varrho/0.917) = C\varrho \tag{11}$$

which by integration over the depth range $z \pm \Delta z$ yields as mean density

$$\bar{\varrho} = (2C\Delta z)^{-1} \log_e [(a - \varrho_{z+\Delta z}/0.917)/(1 - \varrho_{z-\Delta z}/0.917)] \tag{12}$$

where C is a constant determined from the observation.

In the remainder of this paper a number of computed temperature profiles are compared with one another and with observed temperatures. The work should be regarded as an experiment aiming to test the effects of varying different parameters, and the extent to which a transient heat conduction state is adequate to explain the observed temperatures.

3. Results

3.1. *Greenland ice cap*

At Station Centrale near Eismitte, where Sorge in 1930-1 discovered the phenomenon of the negative temperature gradient, Heuberger (1954) has measured ice tem-

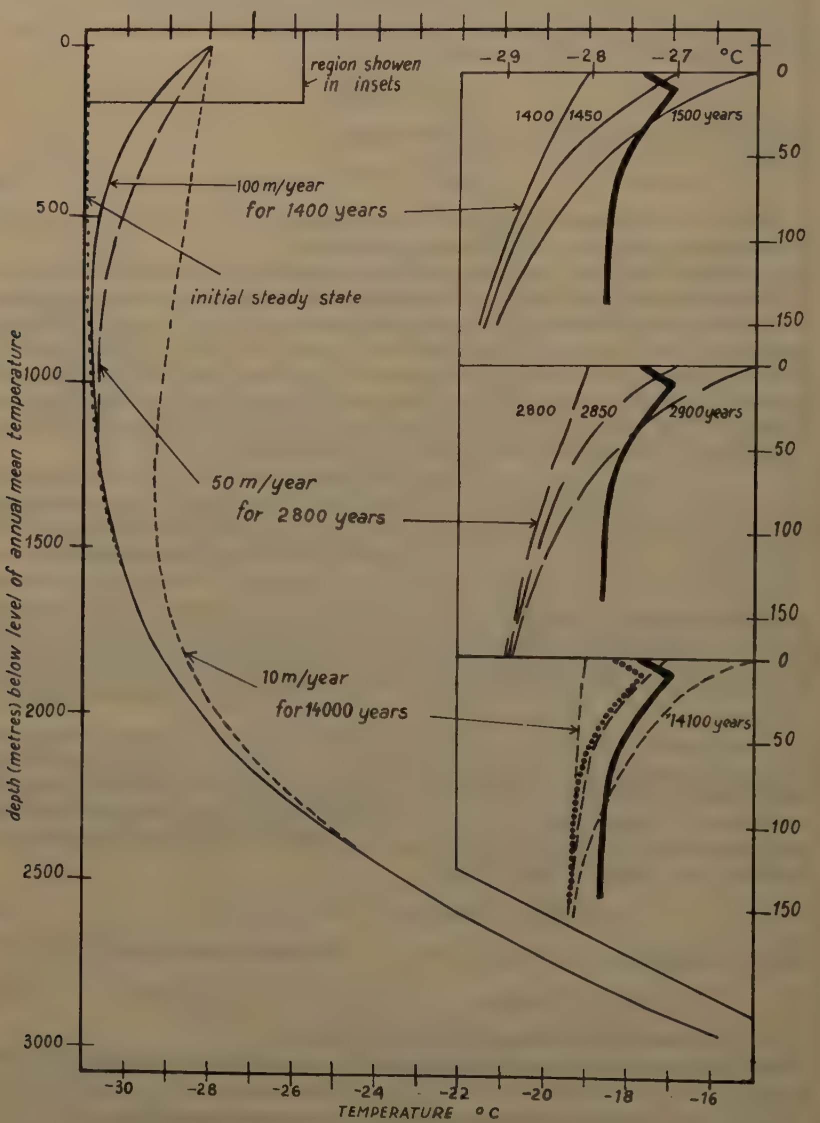

Fig. 2 — Temperature distributions for the Greenland ice cap at Station Centrale. Inset diagrams show the observed temperatures (heavy lines) and the effects of a short-period (100 years) climatic change.

peratures to a depth of 150 metres. Our calculation made the assumption that the ice came originally from the highest point of the ice cap, 140 km away and 200 m higher in elevation. The ice was assumed to have a constant thickness of 3 000 m and to have started moving from a steady state of heat conduction corresponding to a surface temperature of —31°C, a bottom gradient of 1°/50 m, and a net surface accumulation of 20 cm of ice per year. During the travel to Station Centrale the accumulation was increased to 30 cm/year while the temperature rose at a constant rate of 3°/140 km. Following the arrival at Station Centrale the surface warming rate was raised to 2°/100 years, so as to simulate the secular temperature trend observed in the Greenland region in recent times.

Fig. 2 shows the results of these calculations for 3 different speeds, viz. 100 m/year, 50 m/year, and 10 m/year. The observed temperatures, shown as heavy lines in the inset diagrams giving details of the surface layer, are in agreement with motion at the slowest rate and a duration of the climatic trend of between 50 and 100 years. This would imply the lowest temperature at a depth of approximately 1 300 m and virtually steady-state conditions below 2 300 m. The bottom temperature would vary with the value assumed for the geothermal heat flux but should be well below the pressure melting point of — 2°C.

3.2. *Antarctic ice cap*

Ice temperatures have been measured to a depth of 300 m at Byrd station (Bender et. al. 1958). A transient representation for these temperatures has been suggested by

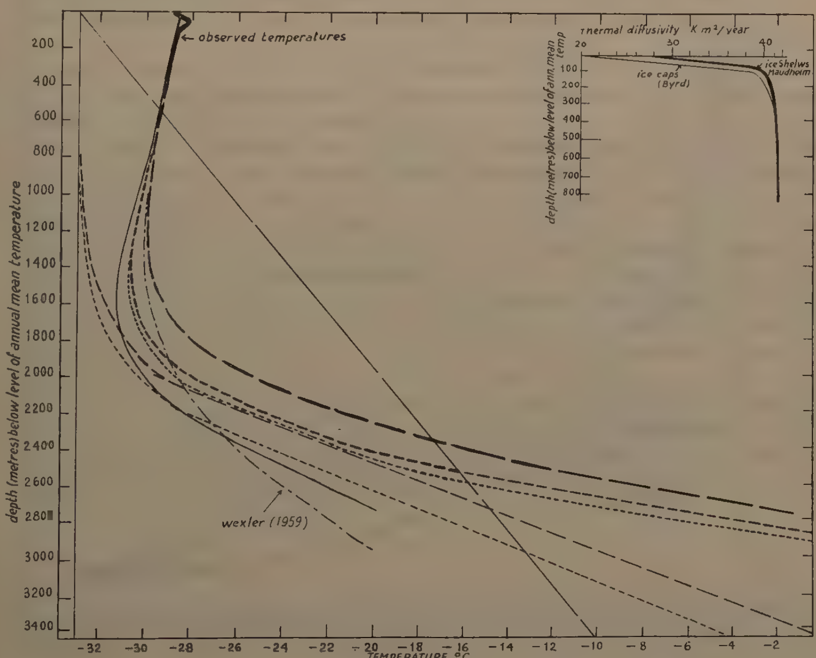

Fig. 3 — Temperature distributions for the Antarctic ice cap at Byrd Station. Thin lines are assumed initial temperature distributions; heavier lines of same type show temperatures after 12,000 years of heat conduction.

Wexler (1959) who assumed that the ice cap was laid down on bare soil at a temperature of — 33°C which then rose steadily by 4°C in 10.000 years during the growth of the ice to its present thickness of 3,000 meters. As an alternative explanation, we have assumed that the ice now at Byrd originated in the highest explored region east of the station where 460 km away the surface rises 680 m above the level at Byrd, giving a mean slope of 1.5×10^{-3}. With a net accumulation of 30 cm/year (Wexler 1959) and the usual surface temperature gradient of 1°/100 m height change an ice velocity of 30 m/year is needed to explain the observed temperature gradient of — 0.15°C/100 cm in the ice by means of (3). Starting with a temperature of — 33°C the ice would reach the observed temperature at Byrd after 12,000 years; during this period it would cover a distance of 360 km and shrink by 540 m, provided the underlying rock surface is level.

Computations have been made for 4 different initial temperature profiles, shown together with the corresponding 12,000 year profiles in fig. 3. In 3 cases the bottom temperature rose above the pressure melting point of — 2°C; the reduction in ice thickness was simulated by cutting off the portion with higher temperatures (without the implication, however, that all or even a substantial part of the shrinkage is due to melting). For the remaining case (which had a constant bottom gradient of 2°/100 m as compared with 4°/100 m for the other cases, the bottom temperature remained well below the melting point when the shrinkage was effected by cutting off one point of the original 70-point grid every thousand years.

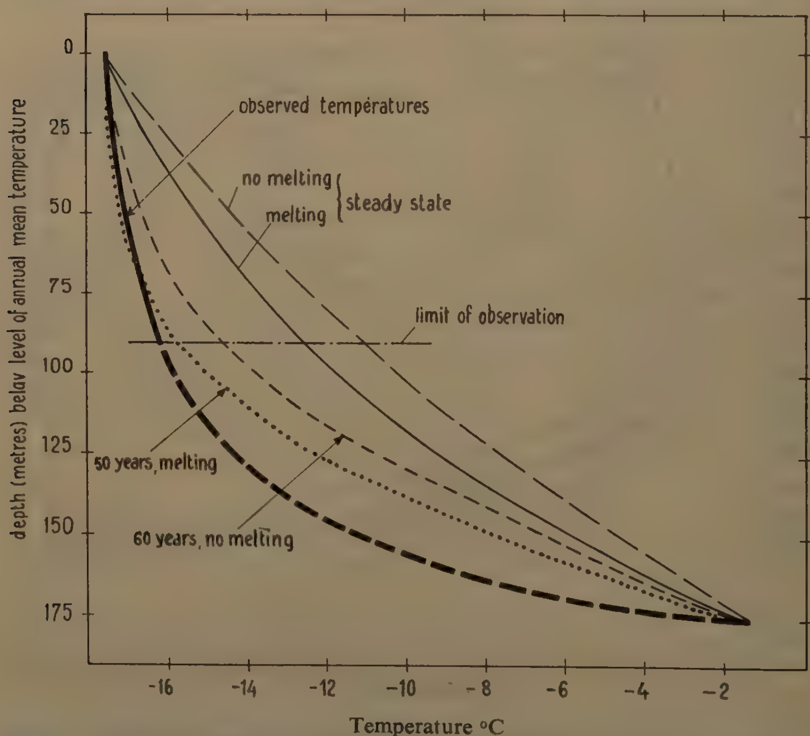

Fig. 4 — Observed and computed temperature distributions for the Maudheim ice shelf.

This discrepancy underlines the need for a more careful consideration of the processes near the bottom of the ice, even though at present the observed temperatures provide no means of distinguishing between different solutions including that of Wexler (1959).

3.3. *Maudheim ice shelf*

Ice temperatures at Maudheim were measured to a depth of 100 m at a distance of some 30 km from the beginning of the ice shelf (Robin 1955). The residence time of the ice in the shelf is probably of the order of 100 years or less. Only two of a number of calculations will be discussed. The results shown in fig. 4 are based on constant surface and bottom temperatures of -17.5 and -1.3°C, respectively; initially the ice was assumed to have the same temperature of -17.5 throughout, except for the lowest point. A distinction was made, following Robin (1955), between the cases of melting, for which the vertical velocity v everywhere equals the net accumulation rate of 44 cm ice/year, and spreading or «no melting» where v decreases linearly to zero at the bottom of the shelf, as for an ice cap. The two corresponding steady-state profiles, obtained after 750 years of conduction, are also shown; despite the use of a variable K they agree very closely with Robin's thoretical curves.

According to fig. 4 the observed temperatures differ little from the melting profile obtained after 60 years of conduction; the «no melting» profile gives already after 40 years a distinctly worse approximation, while the steady-state profiles required intervals an order of magnitude larger than the likely residence time of the ice.

Below 100 m the «observed» temperature profile has evidently been drawn on the basis of indirect evidence of strong melting. No attempt has been made to incorporate this effect or a thickness reduction into the present calculations.

3.4 *Ross ice shelf*

Our final case was based on the temperatures observed in a borehole through the Ross ice shelf near Little America (*).

A number of calculations with suitable surface warming and accumulation rates was made in order to reproduce the observed temperatures on the basis of a constant shelf thickness of 260 m. The results in fig. 5 show the effects of varying the thermal diffusivity, and the other parameters, but all the curves seem quite irreconcilable with the observed profile. Next the profiles in fig. 6 were obtained by allowing for a 17% shrinkage (from 310 to 260 m) in the final 80 km from the ice front (corresponding to 200 year's travel). This was again done by letting the bottom temperature rise without restriction and identifying the actual bottom of the shelf with the -1.3°C isotherm. The dotted curve in fig. 6, which fits the observed temperatures exactly, was obtained by combining a fairly large constant temperature gradient at the original bottom level during the first 175 years with an extremely large one for the remaining 25 years (i.e. near the ice front), together with a moderate amount of surface warming (3° in 200 years or 80 km). This appeared to be the only suitable combination; its implications in terms of heat flux and melting remain to be worked out.

4. Conclusions

The transient solutions here derived for the equation of heat conduction in a moving medium seem to explain a good deal of the observed temperature distributions

(*) We are indebted to Mr. Malcolm Mellor, of SIPRE, for these measurements, taken from the Post Operational Report of IGY project 4.7. They appear to have been published since by Dr. Wexler in J. Glac. March 1960.

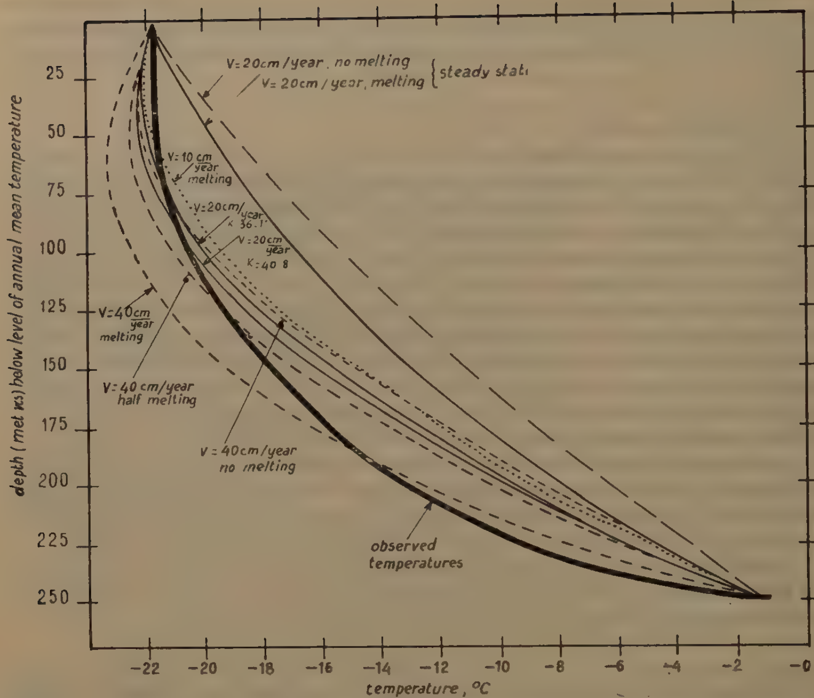

Fig. 5 — Temperature distributions for the Ross ice shelf at Little America, computed for a constant ice thickness of 260 metres and surface warming of 90/80 km.

as due to the ice motion. Climatic trends could produce similar effects only provided they acted over very long periods; experiments with short-term variations in the surface warming rate fully supported Loewe's (1957) conclusion that their effects vanish within an interval comparable to their own duration. On the whole then the persistent thermal effects of ice motion first noted by Robin (1955) are probably paramount.

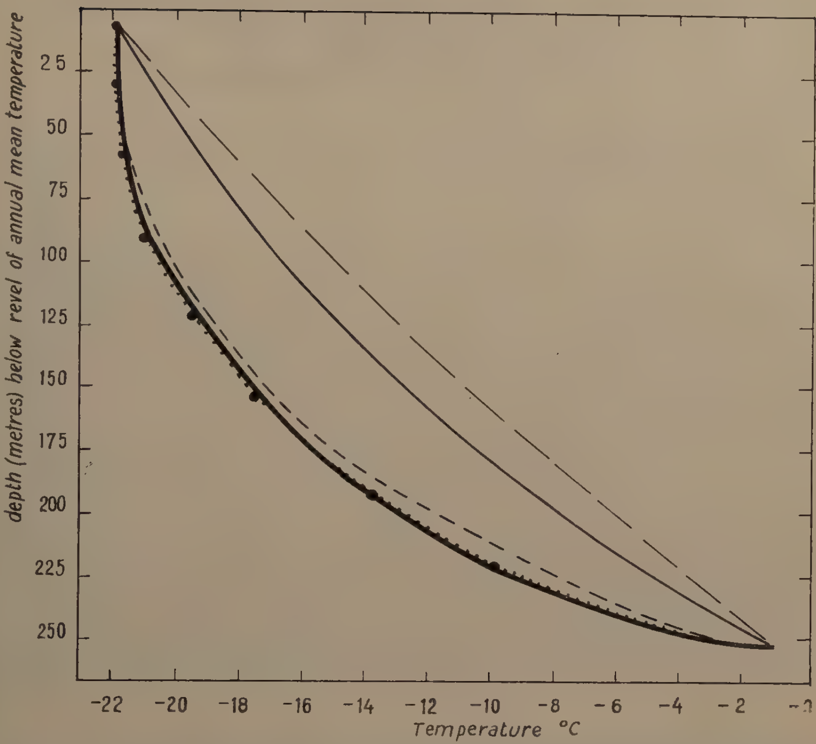

Fig. 6 — Temperature distributions for the Ross ice shelf at Little America, computed for a thickness reduction of 17% and surface warming of 3°C in the final 80 km from the ice front.

observed temperatures
melting { constant thickness, steady state bottom temperature
no melting { – 1.3°C
17% thickness reduction in 200 years; 48‰ m
gradient at original Final 25 years
bottom level 32‰ m for 115 years, 400‰ m

5. Acknowledgement

The writers wish to thank Dr. F. Loewe for his invaluable advice on Arctic and Antarctic conditions, and Miss. J.H. Martin for her help with the numerical results.

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GLACIOLOGICAL RESULTS OF TRAVERSE GEOPHYSICAL OBSERVATIONS IN WEST ANTARCTICA

C.R. BENTLEY

SUMMARY

Data have been collected from traverses covering a total of some 6,000 km on the continental ice of West Antarctica. The geophysical program, including seismic, gravimetric, magnetic, and altimetric observations has produced a considerable amount of glaciological information. The ice is much thicker than had been expected, with a maximum recorded value of 4270 meters. This great thickness is a reflection of the vast channel in the rock floor extending from the Ross Sea toward the Bellingshausen Sea. The ice surface contours on the continental ice show two high areas, one southwest of the Sentinel Mts., the other in the vicinity of the Executive Committee Range. From this and the subglacial topography it is concluded that the ice sheet in West Antarctica was formed by the convergence of two separate ice caps which formed in the two mountainous areas. The Horlick Mountains apparently form an effective block to ice flow from the South Polar Plateau.

A comparison of ice thickness and surface slope from one traverse shows poor correlation, probably due to rough subglaceal topography. An average of about 0.4 bar is obtained for the basal shear stress in an area of little relief. Seismic evidence indicates a layer at the bottom of the ice of about 200 meters thickness, perhaps at the pressure-melting point. Detailed study of the upper snow layers has resulted in an empirical relation between seismic wave velocity and density from which density-depth curves can be drawn and the depth to «pure ice» determined. The variation of Poisson's ratio with depth has been deduced, showing generally a high value near the surface, a minimum at about 10 meters depth, and a gradual increase below toward a value of 0.33. Some evidence of wave velocity variation with direction suggests a possible preferred crystal orientation in the ice.

RESULTS OF THE 1959-60 AIRBORNE TRAVERSE

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SUMMARY

A ski equipped DC-3 aircraft has been used for geophysical studies in Antarctica during the past three years. The scientific program has included geological studies at selected mountain ranges and nunataks, seismic soundings of ice thickness in critical areas, glaciological studies of the snow surface, elevation surveys by barometric and radar altimetry, exploration flights for map making, and weather reconnaissance flights in support of aerial photography by larger aircraft. The gravity work has included measurements between Antarctic staions in order that a unified gravity network might be established on a continental basis. Fourteen hundred kilometers of airborne magnetic profiles have been obtained with a proton precessional magnetometer; from these data it is possible to estimate ice thickness and to infer, in a broad sense, the lithology of the buried rock.

The major objective of the past two seasons has been to resolue the longstanding dispute concerning a possible division of the Antarctic continent by a low lying area between the Ross and Weddell Seas. Landings have been made along two meridians, 90°W and 130°W, in the critical region. The following conclusions are drawn from this work :

1. The topography of the snow surface between the two seas is saddle shaped, with the lowest elevations occurring immediately to the north of the Trans-Antarctic Range. The Antarctic ice cap is therefore not a simple snow dome, but may have been formed by the coalescence of several originally separate ice caps.

2. The ice between the Ross and Weddell Seas is grounded, eliminating the possibility of water interchange at the present time.

3. Most of the rock surface of West Antarctica is far below sea level; the only place where a land connection between East Antarctica and the Palmer Peninsula exists is along meridian 90°W.

4. Even along this meridian, there are places between nunataks where the rock surface plunges below sea level. However, the greatest distance between nunataks is only 150 kilometers, so that no broad trough separates the two seas.

5. The nunataks along meridian 90°W are either granites or strongly deformed sediments and thus differ from the undeformed, block faulted rocks of the Trans-Antarctic Range. The latter range probably continues across the continent to Coats Land on the east side of the Weddell Sea. The nunataks along meridian 90°W are more closely related to the Andean type structures of the Palmer Peninsula.

SUB-ICE TOPOGRAPHY OF ANTARCTICA, LONG. 160° W TO 130° E

A.P. CRARY and F.G. VAN DER HOEVEN

SUMMARY

The data presented are the results of three U.S. traverses in interior Antarctica south and west of the Ross Sea. The Ross Ice Shelf traverse of 1957-58 found that this shelf was floating throughout most of its known area with ice thicknesses 300 to 800 meters, and water depths to 1400 meters below sea level. In 1958-59 the first Victoria Land traverse party travelled up the Skelton Glacier from the Ross Ice Shelf to the plateau and thence westward along the 78°S. Lat to about 131°E. Long. The rock surface underneath the ice dropped below sea level west of the Antarctic horst, then rose at the western and of the line to a few hundred meters above sea level. The second Victoria Land traverse party of 1959-60 proceeded northwest from the top of the Skelton Glacier to the southern point reached by the French traverse parties from Charcot Station, then east to about 73°S Lat and 162°E Long, where the party was flown out by plane. Most of the area covered by this party was found to have rock surfaces below sea level. Combining these results with those given by Imbert of the French expeditions and Pratt of the Trans Antarctic Expedition a clearer picture is obtained of the Atlantic horst, which rises 2000 to 3000 meters above the land surface on either side.

Three traverse operations have been made by U.S. parties on the Ross Ice Shelf and Victoria Land Plateau. The first of these, in the summer of 1957-58, was entirely on floating ice and circled the shelf from Little America to Minna Bluff, to the Beardmore and Liv Glaciers and returned just west of Roosevelt Island. A short fall traverse in March 1958 covered the 200 kilometers of the shelf along the Byrd tractor trail to grounded ice. In the Antarctic summer of 1958-59, the party crossed the ice shelf from Little America, examining the area south of the Minna Bluff-Mount Discovery peninsula, then proceeded up the Skelton Inlet and Glacier to the plateau area, making detailed studies on the glacier en route. On the plateau the traverse route ran due west along the 78° parallel to 131°E, a distance of about 620 kilometers west of the mountains. The return was along the same route back to Minna Bluff, and from there the party turned north and explored the area to McMurdo Station on Ross Island. During the last summer, 1959-60, the traverse again proceeded into the high plateau area along the same Skelton Glacier route, but turned northwest after reaching the plateau, to the furthest south point reached by the French party in 1957-58, approximately 71°S and 139°E. A course was then set along the 72° parallel to the mountains south of Rennick Bay at 162°E, from which the party was eventually airlifted to McMurdo. These three traverses, together with the traverse inland from the Dumont d'Urville Station by the French Expedition and the UK Trans-Antarctic-Expedition whcih, in crossing the continent, travelled from the geographic pole north between the 140 and 160° East Longitudes to the Skelton Glacier, have covered in a very cursory manner, the Antarctic sector between 160°W and 130°E, and allow a preliminary description of the area to be outlined. For the most part, the traverses were of an exploratory nature, and while limited studies of various disciplines were undertaken, the main contributions came from the determinations of the surface elevations, general snow character and annual accumulation, ice thickness values with the aid of the seismograph, gravity and magnetic studies.

1. Surface Elevations

Surface elevations were obtained by altimeters, and in the case of the U.S. traverse parties, a multiple altimetry system was employed, in which differences in elevations between sites 5 to 10 kms apart were determined by differences in simultaneous readings of matched altimeters at the two sites, or by differences in single altimeter readings at the two sites corrected for atmospheric change as read on a stationary second altimeter. Unfortunately, the points of known elevation, so important in any altimetry method, were rarely available except at the beginning and at the end of the traverse, and the accuracy of the values are probably about 20 meters. Figure 1 shows the approximate elevations of the sector under study: the low-lying shelf with elevations 40 to more than 100 meters, and the high plateau with elevations 2200 to 3000 meters above sea level. Several features of these elevations are of glaciological interest. On the Ross Ice Shelf, an increase in elevations is obtained toward the southeast, and this must generally reflect the direction of incoming ice that is responsible for the nourishment of the shelf. The high floating ice elevation directly east of Roosevelt Island is probably the result of crowding from the south between the island and grounded ice to the east. A noticeable drop in elevations was encountered along the Byrd Trail as the crevassed zone that marked the junction of the floating and grounded ice was approached. The significance of this is not apparent from the data available but would seem to indicate that there is little flow of ice from Byrd Land in this area.

The nearly constant elevations along the western portions of the shelf are proof that the high plateau area of Victoria Land contributes little to the shelf. This has been confirmed at least in the Skelton Glacier where the annual flow of ice was found to be only sufficient to take care of the estimated accumulation in the local névé fields.

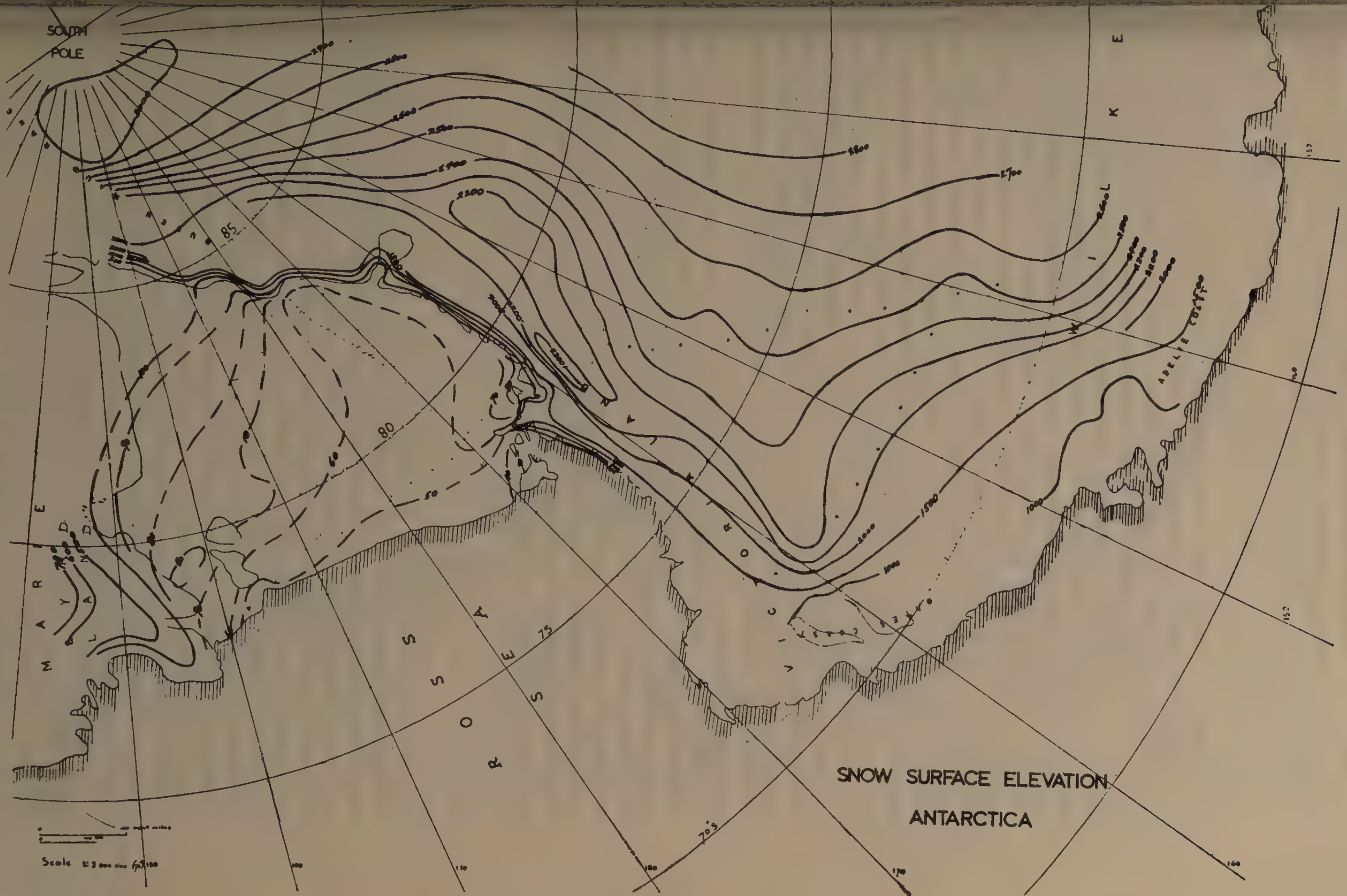
On the plateau the two east-west lines show a general rise to the west of 2 to 3 meters per nautical mile. Immediately west of the mountains along the 78° parallel, however, a drop of 150 meters was found in about 40 nautical miles before the rise began into East Antarctica. This is still another indication of the retreat of the high plateau ice, which is probably the cause of the present «dry» valleys located west of McMurdo Sound. The high area on the Ross Sea side of the Pole, noted by Amundsen and Scott in 1912, was confirmed by the TAE expedition, with indications that the highest part may lie toward Byrd Land. This should represent a buried ridge but proof is not available as yet.

2. Ice Thicknesses

Ice thicknesses in all cases were obtained with the reflection or refraction seismograph method with artificial explosions set up near the surface. Recordings on the high plateau were generally quite poor compared with those obtained on the Ross Ice Shelf, and at only a little more than about one-half of the U. S. sites occupied are reflection depths considered reliable. At the top of the Skelton Glacier and at 78°S, 135°E, seismic refraction results confirmed the reflection thickness values.

On the Ross Ice Shelf the water depth determinations were obtained as well as the thickness values. The reflection from the ice-water interface were much more difficult to obtain than from the water bottom, and generally indirect methods of obtaining ice thickness such as multiple reflections were required. The Ross Ice Shelf ice thicknesses, as seen in Figure 2, are in accordance with the elevations, as would be expected. Anomalies in the ice thickness values are probably due to horizontal forces associated with local grounded ice.

The two east-west lines on the plateau both show several good depth determinations and the ice in both areas was found to be quite thick, the highest values being



SOUTH POLE
SNOW SURFACE ELEVATION
ANTARCTICA
ADELIE COAST

Fig. 1

over 3200 meters. This high ice thickness is in sharp contrast to the low values obtained by the TAE expedition, particularly between 80 and 80°S near 150° East Longitude.

3. Land Elevations

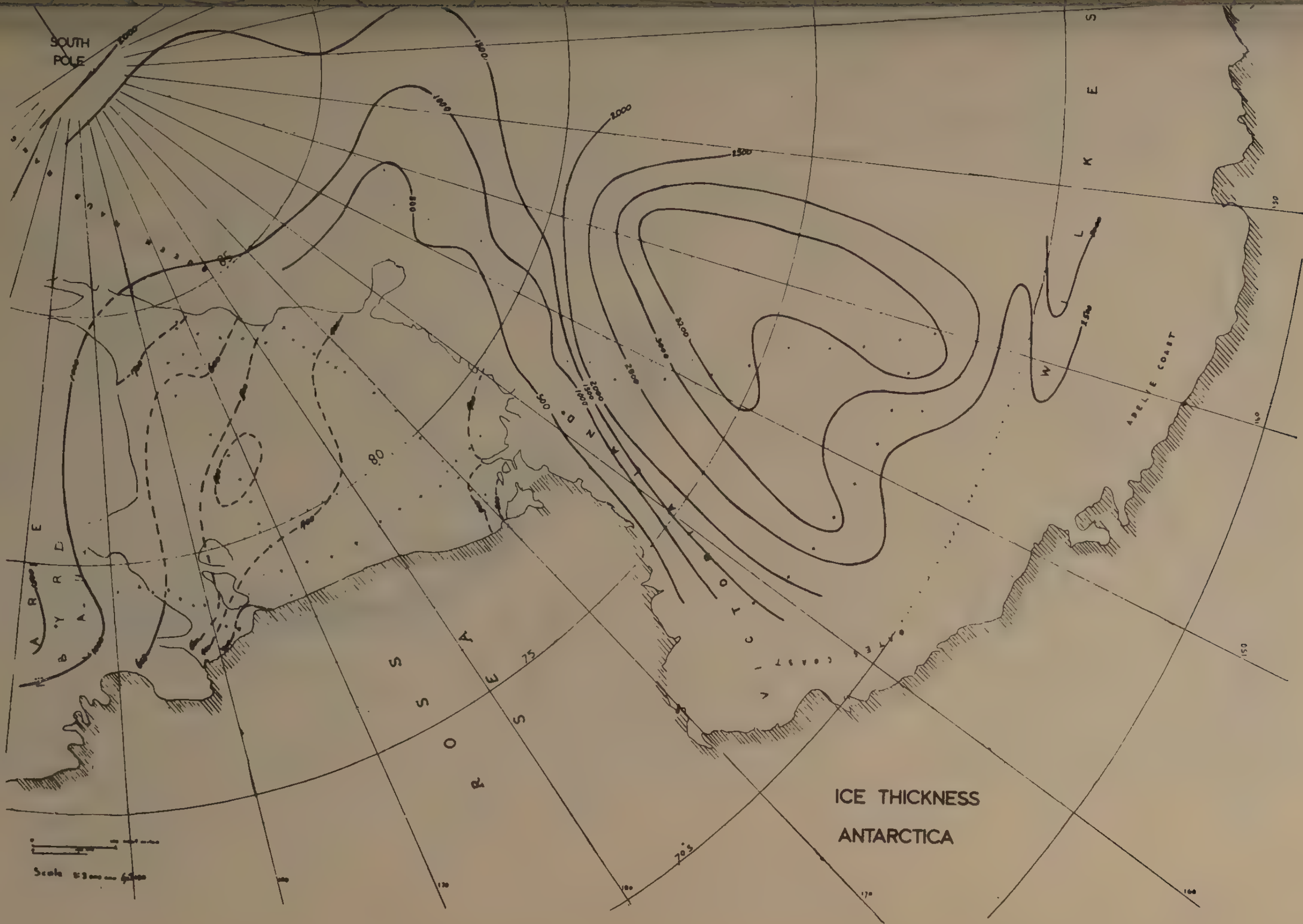
Figure 3 shows the rock surface elevations in the Victoria Land plateau area, as obtained from the two previous charts. Also shown are the ocean bottom levels under the Ross Ice Shelf and in the Ross Sea area, the latter obtained from charts of the U.S. Navy Hydrographic Office. The general appearance of the bathymetry of the Ross Sea including that under the shelf is one of extensive moraines, and presumably the Pennell and Iselin Banks are formed of morainal material. A further high area west of the Pennell Banks is shown in Figure 3, and is a general result of the many additional soundings that have been made in this area since 1957. Along the west coast of the Ross Sea and Ross Ice Shelf. the deepest waters of the continental shelf are located. Deep throughs are present south of the Lady Newness Ice Shelf, and south of Mount Discovery under the Ross Ice Shelf, where a depth of ocean bottom 1400 meters below sea level was found. Such deep troughs in the continental shelves are well developed in the Arctic regions north of USSR, and in the surrounding seas off Greenland and the Canadian Archipelago. They have been attributed to the action of glaciers at a period when the sea surface stood at a lower level. These in the Ross Shelf may also result from downwarping of the crust under the weights of the uplifted Horst area or the volcanics such as Ross Island and Mt. Discovery. With the discovery of the major sub-ice channel between the Ross and Bellingshausen Seas, the Ross Ice Shelf loses its identity as a separate feature and merely forms the western end of this major sub-ice sea.

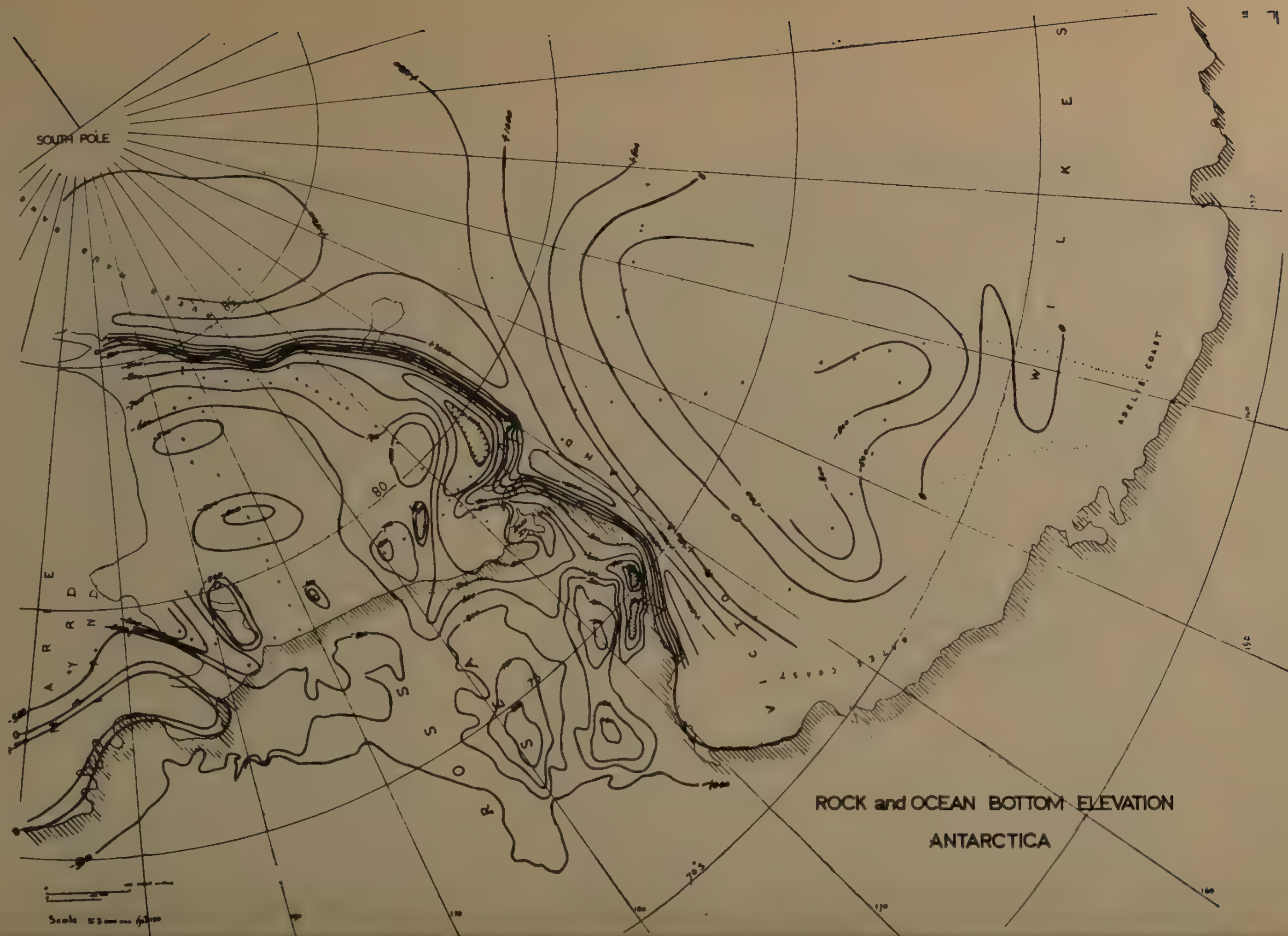
On Victoria Land, the rock elevations drop off to the west, and most of the northern area west of the mountains is below sea level. The suddenness of the drop is not as great on the western side as on the eastern, and at a point nearly 150 kms west of the mountains it is at sea level. The deepest rock elevations lie nearly 400 kms to the west, and even here the rock surface should rise above sea level by isostatic uplift if the ice were removed. South of about 80°S the rock surface is well above sea level, and appears to indicate the presence of a major mountain range nearly at right angles to the Antarctic Horst and running well into the East Antarctic continent.

4. Gravity Observations and Refraction Results

Values of free-air gravity anomalies on the high plateau area along the 78°S parallel vary from + 50 to −55 milligals, averaging −26 milligals. These free-air anomalies show good correlation with the rock depths determined from seismic results; the rock is above sea level at either end of the line where gravity free-air anomalies are positive, and the rock surface is below sea level where gravity free-air anomalies are negative. The area appears generally to be in approximate isostatic equilibrium.

At the Little America Station on the eastern side of the Ross Ice Shelf a reversed refraction profile gave a 4.24 km/sec velocity layer at a depth of 1170 meters below ocean bottom, overlain presumably by sediments. In the McMurdo area one unreversed profile near the Hut Point Peninsula showed a higher velocity, 5.1 km/sec at a depth of 1700 while another between Hut Point Peninsula and White Island, also unreversed, and of poorer quality indicates a velocity of 4.1 km/sec with over 3 kilometers of sediments. Undoubtedly the shallower area of the Ross Sea and shelf area sompared to the channel under the Byrd Land ice is the result of glacial deposition of sediments.


ICE THICKNESS
ANTARCTICA
SOUTH POLE
ADELIE COAST
OATES COAST
VICTORIA LAND
WILKES
BYRD
ROSS SEA
2000
1500
1000
500
800
2500
2200
3000
2800
80
75
70°S
150
160
170
180
Scale 1:3 000 000



SOUTH POLE
WILKES
ADELIE COAST
OATES COAST
VICTORIA LAND
ROSS SEA
EDWARD VII
80
70°S
ROCK and OCEAN BOTTOM ELEVATION
ANTARCTICA

At the top of the Skelton Glacier névé field a refraction profile showed a 600 m thick layer with a velocity of 4.3 km/sec under the ice. This was underlain by a layer with a velocity of 5.43 km/sec. Farther along the same parallel near 135°E a reversed profile showed a velocity of 5.82 km/sec directly under the ice, with no indication of the lower, and presumably Beacon Sandstone formation.

DEEP DRILLING IN ANTARCTICA

J.A. BENDER[1] and A.J. GOW[2]

SUMMARY

A modified rotary well drilling rig with compressed air as the drilling fluid was used successfully in the Antarctic to drill in snow and ice. Excellent cores were obtained down to 308 meters at Byrd Station and down to 254 meters in shelf ice at Little America V. Detailed analysis of the cores indicates the annual accumulation to be about 15 cms of water equivalent at Byrd Station and 21 cms at Little America V. Depth density and depth-temperature curves for the two stations are presented. Temperature, closure and inclination measurements have been continued on the hole at Byrd Station. These indicate that over the past two years, the temperature with depth has not changed, that there has been no inclination, and that the drill hole si closing very rapidly with depth, and non-linearly with time.

1. Introduction

Glaciers have long been considered as indicators of climate, but most of the older investigations were confined to temperate glaciers because of the inaccessibility of the polar areas. Approximately two-thirds of the area of the Greenland ice sheet and most of the Antarctic ice sheet are considered as «dry», i.e., all precipitation is snow with little or no melting. Thus, each year's precipitation and everything that fell with it or on it are preserved and buried under the following year's snowfall. In more recent years, access to the polar areas has provided the opportunity to investigate past events, literally frozen in time, through the dating and analysis of snow and ice samples obtained at depth.

The United States IGY Deep Drilling Program was assigned to the U.S. Army Snow Ice and Permafrost Research Establishment (USA SIPRE), Corps of Engineers.

Drilling equipment and techniques were tested and developed by USA SIPRE during the summer of 1956 in Greenland, where a hole was drilled to about 300 meters. Another hole was drilled to 400 meters in Greenland during the 1957 summer and good core recovery made.

A second drill rig and equipment were shipped to the Antarctic during 1957 and transported from Little America V to Byrd Station by sled-train in October, 1957. Most of the glaciological equipment and accessory drilling parts were parachuted from C-124 Globemasters in December, 1957. Drilling started in December, and was completed by the end of January, 1958 (Technical Report 60). Excellent cores were obtained between 20 and 300 meters below the surface. The drilling equipment was then returned by sled-train to Little America V, in preparation for the next year's drilling program. Drilling operations started on the Ross Ice Shelf during the latter part of October and terminated during the middle of December, 1958 (Technical Report 70). Total depth drilled was 254.2 meters and the total thickness of the Ice Shelf at this location was estimated between 256 and 259 meters. The 308 meter Byrd Station hole was remeasured in December, 1958, and January, 1960.

2. Drilling techniques and operations

A modified rotary well-drilling rig was mounted on a 10-ton sled and compressed air used as the drilling fluid. The heated, compressed air was cooled to within 2-3°C

(1) U.S. Army Snow Ice and Permafrost Research Establishment
(2) Arctic Institute of North America

of the ambient air temperature by an air-to-air heat exchanger. A special core feed device was used to insure a uniform rate of drilling. Drilling bits cut a 37/8-inch (9,8 cm) cose and a 5 3/4 inch (14.6 cm)hole. It was necessary to case the upper portion of the drill hole to prevent the dissipation of the compressed air into the permeable upper snow. At Little America Station, the last 5.5 meters were drilled, using diesel oil as the drilling fluid to counteract a possible inrush of sea water as the shelf bottom was peirced. A marked improvement in the condition of the core was observed.

The ice cores were examined in transmitted light for changes in grain size and density. Stratigraphy consisted of alternating layer of coarse and fine-grained snows. The coarse-grained layers are considered to have originated under summer conditions, and the finer grained, denser layers were interpreted as winter deposits. At Little America, the summer layers were often associated with ice lenses, ice pellets, and other obvious melt water features. No signs of melting were observed in the cores from Byrd Station. Stratigraphy was carefully recorded and the cores then cut at natural stratigraphic breaks for density determinations. Distinguishable stratigraphy can be observed down to about 120 meters. Below this, the material becomes too homoge-

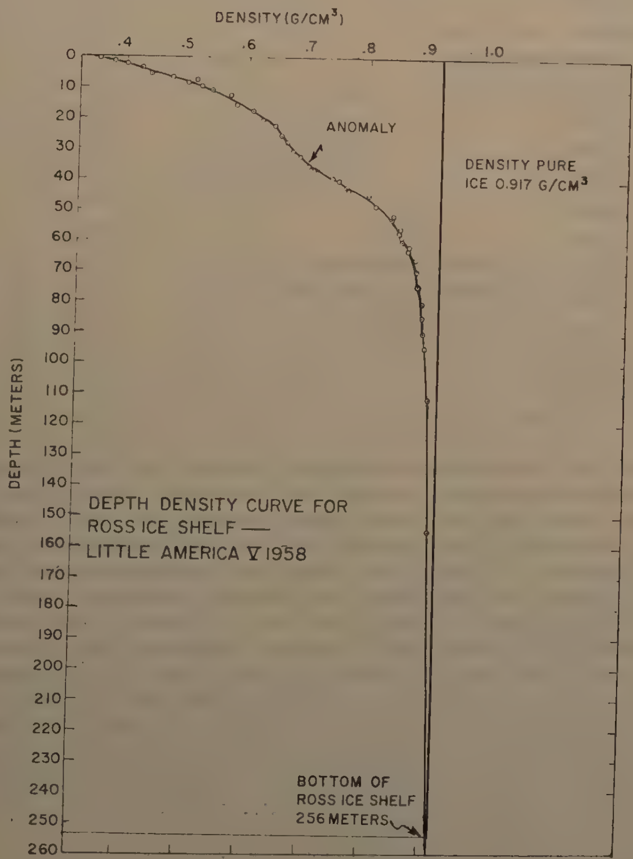


Fig. 1 — Depth density curve at Little America V.

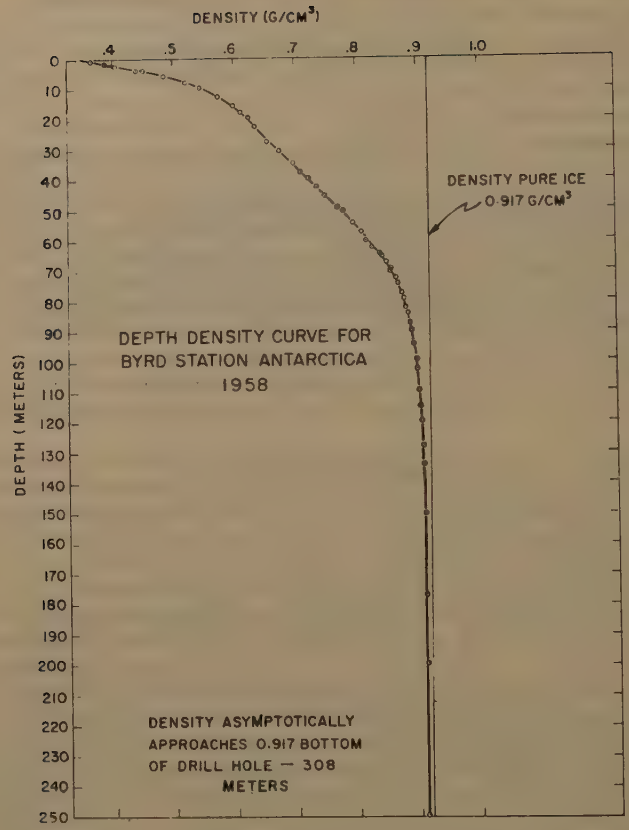


Fig. 2 — Depth density curve at Byrd Station.

neous, and other techniques are required to determine the annual accumulation.

To aid in the glaciological investigations, shallow pits were excavated near the drilling sites at both Byrd and Little America Stations, and detailed snow studies were made. Three-inch (7.7 cm) diameter cores were also obtained with a hand auger down to a depth of 21.6 meters at Byrd Station and 18.0 meters at Little America.

Depth density curves for the two sites are-shown in Figures 1 and 2. Table 1 represents the accumulation record at Byrd Station plotted in terms of the 5-year means and the annual running mean. This phase of the work has not been completed for the Ross Ice Shelf cores, but preliminary studies on an 18 meter section of hand augured core indicated that 21.0 cm of water-equivalent/year is a fair value for the average annual rate of accumulation. If it can be assumed that the Ice Shelf, in the vicinity of the drill site, is composed entirely of depositional strata (melting at the bottom having eliminated the original basement of the shelf), then on the basis of a constant annual accumulation of 21 cm of water, it may be said that Little America V rests upon approximately 1,225 years of accumulated snow.

Most of the cores were split lengthwise and one-half transported under refrigeration back to the USA SIPRE Laboratory in Wilmette, Illinois. Further research at

the laboratory will include the detailed completion of the density profile, grain size and fabric studies, analysis of particulates and soluble materials, measurement of gas bubble pressure, and chemical analysis of entrapped gas. Measurement of the O^{18}/O^{16} ratios is presently being carried out at the California Institute of Technology.

3. Temperature measurements

Temperature measurements were made in the Byrd Station deep hole three days

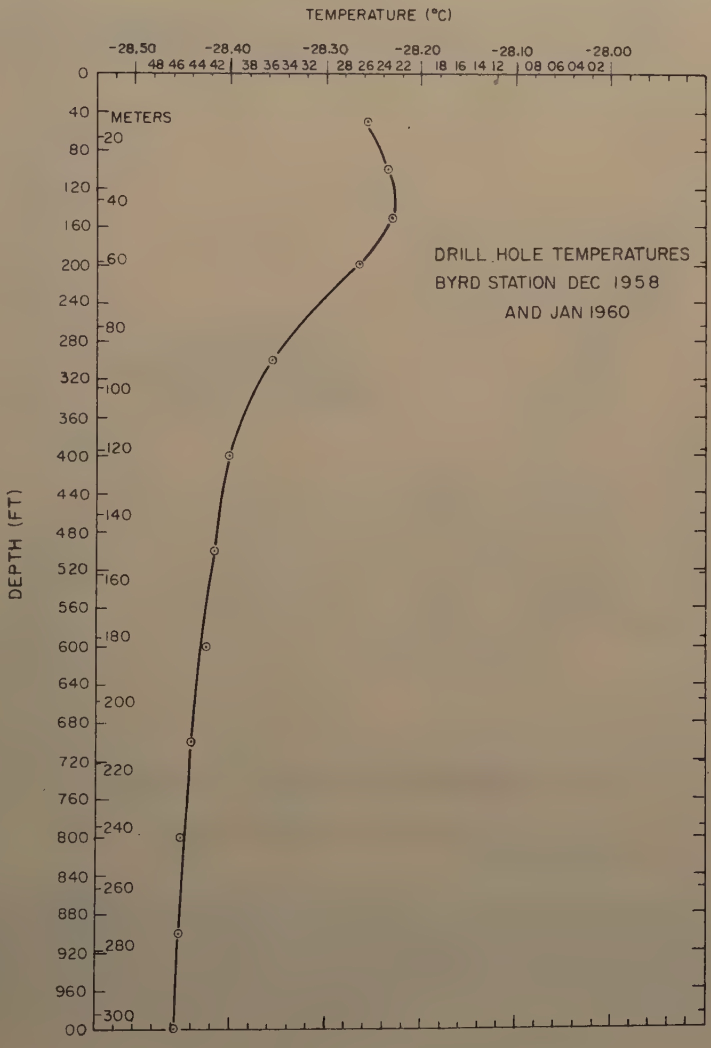


Fig. 3 — Temperature with depth at Byrd Station.

after completion of the drilling. A 100-ohm copper, four-wire thermohm with a Wheatstone bridge temperature indicator was used to make the measurements. A specially designed temperature probe with four thermistors and a platinum resistance thermometer was used for measurements in both the Little America and Byrd deep holes in December, 1958, and remeasurement of the Byrd hole with the same probe was carried out again in January, 1960. The depth-temperature curves are shown in Figures 3 and 4. It may be noted that there has been no change in the temperature during the past year in the Byrd Station hole.

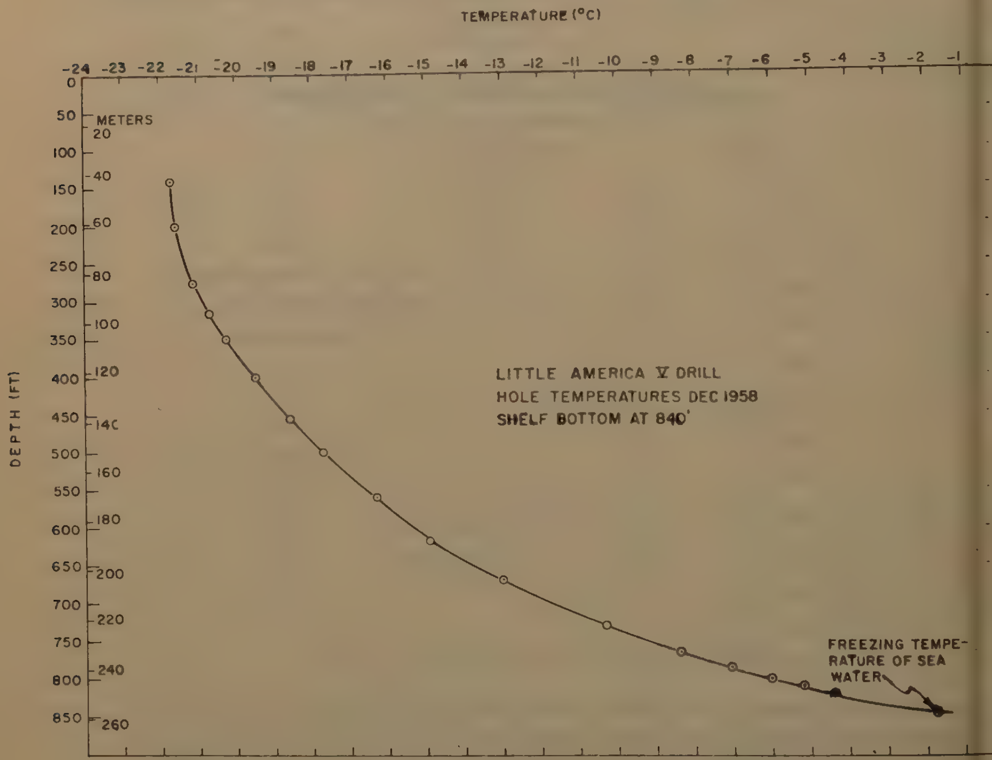


Fig. 4 — Temperature with depth at Little America V.

4. Hole diameter measurements

The diameter of the Byrd drill hole at various depths was measured in December, 1958, and again in January, 1960. Measurements were made with a specially designed electric caliper which varies a resistance as a function of the diameter. The depth-diameter and load-diameter curves for the two years of measurements are shown in Figures 5 and 6.

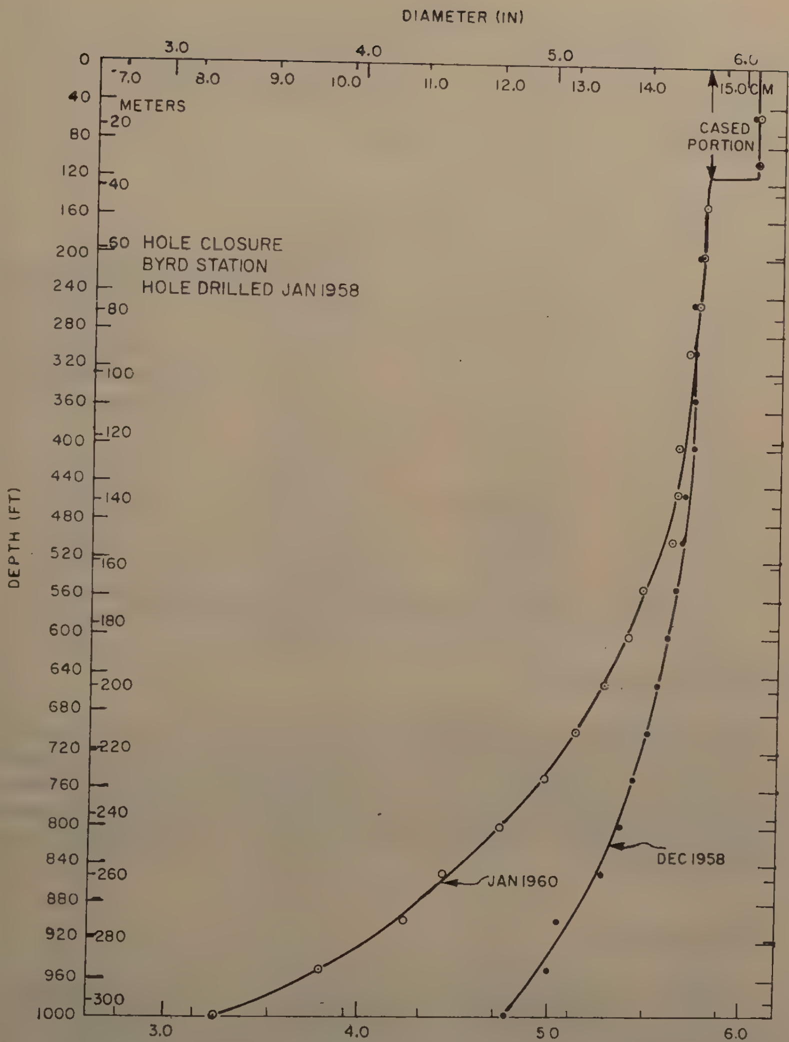


Fig. 5 — Drill hole closure with depth at Byrd Station.

5. Hole inclination measurements

Hole inclination measurements have been made in the Byrd Station hole, but, to date, only very small changes have been detected. These are still within the error of measurement, and are considered to be insignificant.

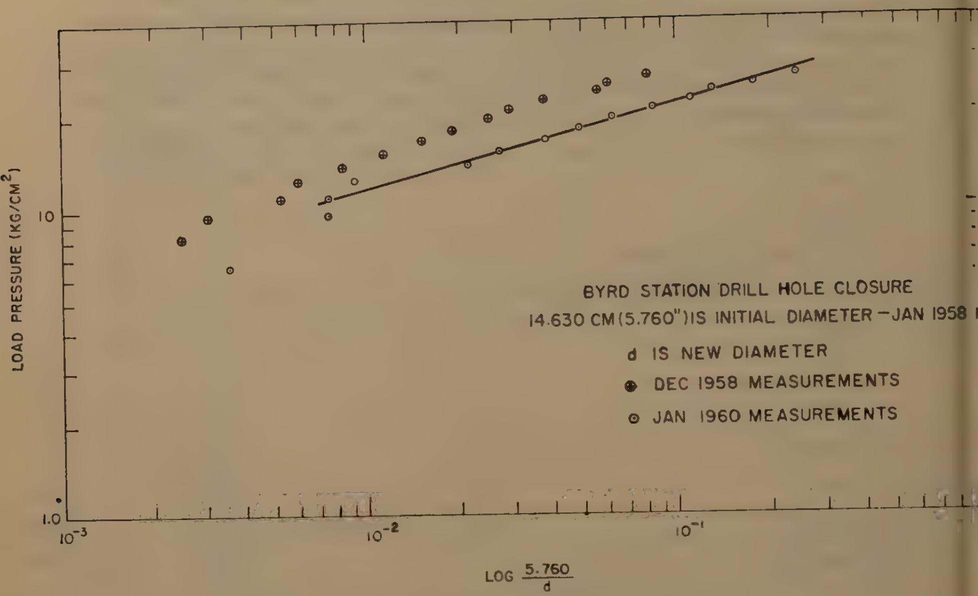


Fig. 6 — Drill hole closure with load at Byrd Station.

6. Future investigations

Work will continue on the laboratory examination of the ice cores, and it is planned to remeasure yearly the temperature, closure and inclination of the drill hole at Byrd Station. A thermal drill rig is currently under development at USA SIPRE to drill and obtain cores down to 3.000 meters. A prototype will be tested in Greenland during the fall of 1960 and later used in Antarctica. This work is under the direction of Mr. B.L. Hansen.

7. Acknowledgments

The Deep Drilling Program has been a combined team effort, and has received support from the United States IGY and National Science Foundation, with logistic support from the U.S. Army, Navy and Air Force. The development of the drilling equipment and techniques in Greenland was under the supervision of G.R. Lange. E.W. Marshall was field leader during the Byrd Station drilling, with Glaciologist, A.J. Gow, Drilling Engineer, R.W. Patenaude, and Chief Driller, J.V. Tedrow. R.H. Ragle was field leader at the Little America site, with Glaciologist A.J. Gow, Drilling Engineer, R.W. Patenaude, and Chief Driller J.V. Tedrow. B.L. Hansen was in charge of the hole instrumentation design, and aided in the measurements. A.J. Gow made the recent remeasurements at Byrd Station. Mr. E.W. Marshall was IGY Project Leader during the first portion of the project, and Mr. J.A. Bender, during the remainder. The entire project was aided and encouraged by Dr. H. Bader.

TABLE 1

Snow Accumulation Record, Byrd Station, from 1958-1547

Years	Depth (meters)	5 Year Mean Accum (g/cm^2)	Running (g/cm^2) Mean Accum
1958-1953	0 - 2.02	14.8	14.8
1953-1948	2.02- 3.70	13.1	14.0
1948-1943	3.70- 5.00	12.2	13.4
1943-1938	5.00- 6.46	14.3	13.6
1938-1933	6.46- 8.05	16.6	14.2
1933-1928	8.05-9.48	15.4	14.4
1928-1923	9.48-11.00	16.8	14.8
1923-1918	11.00-12.20	13.7	14.6
1918-1913	12.20-13.57	15.7	14.7
1913-1908	13.57-14.73	13.8	14.6
1908-1903	14.73-15.96	14.9	14.6
1903-1898	15.96-17.12	13.9	14.6
1898-1893	17.12-1811.	12.0	14.4
1893-1888	18.11-19.45	16.5	14.5
1888-1883	19.45-20.63	14.9	14.7
1883-1878	20.63-21.92	16.2	14.7
1878-1876	21.92-22.44	13.8	
TRANSITION FROM AUGER CORES TO MAIN DRILL CORE			
1876-1871	22.44-23.62	14.9	14.9
1871-1866	23.62-24.80	16.1	14.7
1866-1861	24.80-25.90	13.7	14.7
1861-1856	25.90-27.16	16.5	14.7
1856-1851	27.16-28.26	14.5	14.8
1851-1846	28.26-29.36	14.7	14.8
1846-1841	29.36-30.34	13.2	14.7
1841-1836	30.34-31.62	17.5	14.8
1836-1831	31.62-32.80	16.2	14.9
1831-1826	32.80-33.98	16.4	14.9
1826-1821	33.98-34-94	13.2	14.9
1821-1816	34.94-36.13	16.8	15.0
1816-1811	36.13-37.31	16.8	15.1
1811-1806	37.31-38.53	17.5	15.1
1806-1801	38.53-39.65	16.2	15.1
1801-1796	39.65-40.67	14.8	15.1
1791-1791	40.67-41.76	16.0	15.1
1791-1786	41.76-42.71	14.0	15.1
1786-1781	42.71-43.82	16.5	15.1
1781-1776	43.82-44.91	16.2	15.2
1776-1771	44.91-46.07	17.4	15.2

TABLE 1 (Cont'd)

Years	Depth (meters)	5 Year Mean Accum (g/cm^2)	Running Mean Accum (g/cm^2)
1771-1766	46.07-47.30	18.6	15.3
1766-1761	47.30-48.29	15.2	15.3
1761-1756	48.29-49.30	15.6	15.3
1756-1751	49.30-50.30	15.5	15.3
1751-1746	50.30-51.39	17.0	
1746-1741	51.39-52.51	17.3	15.4
1741-1736	52.51-53.50	15.6	
1736-1731	53.50-54.38	14.0	15.4
1731-1726	54.38-55.47	17.3	
1726-1721	55.47-56.43	15.3	15.4
1721-1716	56.43-57.22	12.7	
1716-1711	57.22-58.22	15.5	15.4
1711-1706	58.22-59.20	16.0	
1706-1701	59.20-60.17	15.8	15.3
1701-1696	60.17-61.15	16.0	
1696-1691	61.15-62.11	15.4	15.4
1691-1686	62.11-63.07	15.8	
1686-1681	63.07-64.05	16.3	15.4
1681-1676	64.05-64.93	15.0	
1676-1671	64.93-65.91	16.5	15.4
1671-1666	65.91-66.80	15.4	
1666-1661	66.80-68.03	20.8	15.5
1661-1656	68.03-68.87	14.3	
1656-1651	68.87-69.77	15.4	15.5
1651-1646	69.77-70.76	16.9	
1646-1641	70.76-71.72	16.2	15.5
1641-1636	71.72-72.75	17.8	
1636-1631	72.75-73.68	16.1	15.6
1631-1626	73.68-74.60	15.9	
1626-1621	74.60-75.41	14.1	15.6
1621-1616	75.41-76.28	15.2	
1616-1611	76.28-77.14	15.0	15.5
1611-1606	77.14-78.16	17.8	
1606-1601	78.16-79.19	17.0	15.6
1601-1596	79.19-80.03	14.7	
1596-1591	80.03-80.88	14.9	15.6
1591-1586	80.88-81.72	14.8	
1586-1581	81.72-82.70	17.3	15.6
1581-1576	82.70-83.62	16.3	
1576-1571	83.62-84.60	16.8	15.6
1571-1566	84.60-85.44	14.8	
1566-1561	85.44-86.36	16.3	15.6
1561-1556	86.36-87.26	16.0	
1556-1551	87.26-88.08	14.6	15.6
1551-1547	88.08-88.78	14.9	15.7

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ON THE THEORY OF GLACIER MOTION

P.A. SHUMSKIY

SUMMARY

All the existing dynamic theories of glacier motion are unsatisfactory for the reason that, being based on the calculation of laminar flow, they cannot properly account for the most important type of movement – block sliding along the bed or along bottom thrust—faults. The calculation of uniaxial longitudinal stresses and deformations supplemented by the calculation of laminar flow should serve as a calculation scheme of glacier motion for the first approximation. The proposed solution for a moving force in a glacier of arbitrary shape under conditions of a plane problem takes into account the density of the ice, the thickness of a glacier, the inclination of a surface and a bed, and the curvature ot the latter. The solution for normal and tangential stresses in a glacier takes into account the distribution of the moving force and the force of friction against the bed in a longitudinal profile of a glacier. The solution for the rates of ice deformation is based on a power relationship between the stress and the strain rate. The solution for the average velocity of movement of a glacier is based on integration of the average rates of uniaxial longitudinal deformation along a glacier. The laminar component of the velocity of movement is determined by means of integrating the rates of shear along the vertical profile of a glacier. The solution for the vertical component of the velocity of movement is obtained by integration of the rate of the uniaxial vertical deformation along the vertical profile of a glacier.

The proposed solutions enable one, for the first time, to determine the distribution of the coefficient of friction against the bed of a glacier from measurements of its shape and movement. Available data show a variability of the coefficient of friction of glaciers against the bed in space and in time.

The propounded theory gives a far more complicated dependence of the equilibrium shape and dimensions of a stationary glacier upon the intensity of accumulation and ablation, the relief of the bed, and the force of external friction, than was hypothesized by previous theories.

The theory permits establishing, separately, the values of run-off (conditioned by the inclination of the bed) and of spreading-out (conditioned by the intensity of accumulation) stresses, deformations, and movement, and thus to analyze the role of the relief and climatic conditions in the dynamics of glaciers.

1. INTRODUCTION

This paper sets out the basic principles of a dynamic theory of the movement of glaciers, worked out by the author, that accounts for the principal types of stresses and strains in ice masses and explains both the most important elements of their motion: block sliding along their bed and laminar flow.

In contrast to the balance kinematic theories, the dynamic theories establish a relationship between the motion of ice masses and the forces operating on them. This problem may be divided into three parts:

1) a determination of the stressed state of the ice from its physico-mechanical properties, the stress of the gravitational force, and the size and shape of the glacier;

2) a determination of the strain rates of the ice from the characteristics of the stressed state and the mechanical properties of the ice;

3) a determination of the velocity of glacier motion by summing the absolute strain rates of the constituent ice.

A consideration of the state of elaboration of the dynamic theories of motion shows that the greatest barriers to a solution of the problem of glacial movement

now are the first and the third points mentioned above. Just recently the resolution of the second problem has made considerable progress, and accounts for the main achievements in the development of dynamic theories of glacier motion. However, it is apparently impossible to solve the first and the third problems that is to integrate the complete systems of differential equations of the theory of deformed mediums in stresses or in velocities of displacements for a glacier. This circumstance compels, when solving the first and second problems, to confine to a consideration of simple stresses and strains, wich to a certain approximation reflect the real conditions.

A solution, in principle, for the velocity of motion of a glacier based on assumptions that the ice experiences only the pure shear strain and is an ideally viscous material, was given in 1921 by C. Somigliana. After J. Glen established the power relationship between stress and strain rate of ice, J.F. Nye modified this solution for conditions of viscous-plastic flow (1952). Connecting the dynamic equation of viscous-plastic flow with the equation of the mass balance, S.S. Vyalov (1958) applied this theory to determine the equilibrium shape of the surface of a glacier (i.e., a glacier in a steady state) resting on a horizontal bed. J.F. Nye (1959) utilized this same power law for a calculation not only of the laminar flow but also of sliding of the glacier along its bed, for a determination of the equilibrium form of the glacier with an account of sliding along the bed and of irregularities in the latter, and also (on the basis of J. Weertman's method (1958) for an analysis of the evolution of deviations of the glacial surface from the equilibrium form.

However, all this progress is based on a prerequisite of calculation of motion only on the basis of stresses and deformations of pure shear, that is too far removed from reality and greatly reduces the applicability of the conclusions obtained. In fact, the underlying calculation scheme is fundamentally unsuited either for a determination of the equilibrium form of the glacier, or for a calculation of the velocity of sliding along the bed, because:

1) laminar flow occurs only in an infinite plane-parallel plate of ice lying on a uniformly inclined basis, and any alteration of the bed inclination or deviation of the surface from the parallelism to the bed is associated with an impairment of laminability;

2) the velocity of glacier sliding along the bed is not so much due to the magnitude of tangential stress, as due to the rate of absolute longitudinal strains of ice in the direction of movement between the fixed beginning of the glacier and a given point.

Further progress in the elaboration of a quantitative theory of glacier movement is possible only on the basis of a more refined system of calculation, which would permit to take into account not only the shear stresses and strains, but also the normal stresses and the strains caused by them.

One of the possible variants is the precise solution of J.F Nye (1957) for distribution of velocity in vertical profile of a glacier where side by side with shear parallel to the bed single-axis longitudinal strain is observed. This solution implies that the velocity on the surface is known, gradient of its change in the direction of motion known and constant, and the longitudinal strain is plane. All this reduces the possibility of applying this calculation scheme.

In the solution given below the author proceeds from the assumption, that it is possible to divide a complex stressed state into two independent systems of stresses, and to determine their sums according to the principle of superposition.

The necessary prerequisite for solution of the problem is the knowledge of connection between stresses and ice strains.

2. THE FLOW LAW OF ICE

As numerous investigations have shown, polycrystalline ice is a viscous-plastic body that flows under the effects of the very slightest stress. The rate of flow of ice is a power function of the stress that it experiences; and the parameters of this power relationship are conditioned by the temperature and structure of the ice, the magnitude of the stress, and, apparently, to some extent, also the magnitude of the hydrostatic pressure. Due to alterations in these conditions, the fluidity of ice in different parts of the glacier differs; however the alterations themselves in the conditions inside a glacier are not sufficiently well known, and total account of them would make the problem of the calculation of motion insolvable. For this reason, without delving into the physical essence of the processes of ice straining, we shall consider only the effects of the factors enumerated above, as far as this is necessary for the choice of the values of averaged parameters of the flow law applicable to certain conditions.

The fluidity of ice is greatly reduced with decreasing temperature. As the experiments of N. Royen (1922) and K.F. Voitkovskiy (1959) have shown, the rate of flow of ice is inversely proportional to the coefficient $\gamma-(\theta-1)$, where θ is the temperature of the ice in degrees centigrade. The fluidity increases sharply with decreasing density, attaining in snow values of $10^4 - 10^6$ times greater than in ice. To a high degree it depends likewise on the crystallographic orientation of ice grains, and, for sufficient stresses, on the size of grains. However, the very structure of the ice changes in the process of its straining, adapting itself to the nature and magnitude of stresses, which fact permits—to a first approximation for the glacier as a whole—to disregard it as an independent factor.

The effect of change in the hydrostatic pressure on the fluidity of ice has not been sufficiently ascertained. Yet it is obvious that an increase in it does not enhance the fluidity, and for a high strain rate (as the experiments of S.S. Vyalov have shown) it even reduces it noticeably. However, at present there are no data permitting an account of this effect.

The power law of flow in the case of a complex stressed state can be expressed in terms of:

$$\dot{\gamma}_i = 2\left(\frac{\tau_i}{k}\right)^n, \tag{1}$$

where $\dot{\gamma}_i$ is the intensity of shear strain, and τ_i — the intensity of shear stresses. This relationship between the principal stresses $[\pm\tau, 0, \mp\tau]$ and strain rates $[\pm\frac{1}{2}\dot{\gamma}, 0, \mp\frac{1}{2}\dot{\gamma}]$ is true for the pure shear (the main components of the stress deviator $\tau' = \tau$). For single-axis stress $[\pm\sigma, 0, \mp\sigma]$ and plane straining $[\pm\dot{\varepsilon}, 0, \mp\dot{\varepsilon}]$ this relationship appears the form:

$$\dot{\varepsilon} = \left(\frac{\sigma'}{k}\right)^n \tag{2}$$

where the chief stress deviators σ' are:

$$\sigma'_x = -\sigma'_z = \frac{1}{2}(\sigma_x - \sigma_z), \quad \sigma'_y = 0 \tag{3}$$

Into (1), and (2) we substitute the absolute values of τ_i, τ and σ', but the $\dot{\gamma}$ and $\dot{\varepsilon}$ obtained are considered positive in the case of tension and shear in the direction of the coordinate axis, and negative in the case of compression and shear in the reverse direction.

The exponent n in the equations (1) — (2) ranged from 1.5 to 4.2 in different determinations, increasing on the average with the growth in stress. However, there is observed a considerable spread in the results. Thus, from data on the rate of compression of tunnels in the glaciers of Skauthöe (Norway) and Z'Mutt (Alps), J.F. Nye obtained $n = 3.07$ for a mean stress $\tau \simeq 0.7$ kg cm^{-2}, whereas in certain experiments carried out by K.F. Voytkovskiy, the mean stress $\tau \simeq 2$ kg cm^{-2} corresponded to $n = 2.2$. An averaging of the results of all available to the author determinations of the value of n by means of laboratory experiments and measurements of the rate of deformation of tunnels and boreholes in glaciers within the limits of stress, under the action of which glacier flow occurs, permits to take $n \simeq \sqrt{3}$ for calculations.

The spread of values for the constant of proportionality k in different determinations is far greater than for n, but this spread is considerably reduced when recalculating with a reduction to the common exponent $n = 3$ and temperature $\theta_0 = 0$°C, on the basis of the requirement of the equality:

$$\dot{\varepsilon} = -\frac{\bar{\tau}^3}{(\bar{\theta}_0 - 1)\,k_0^3} = -\frac{\bar{\tau}^{n_1}}{(\theta_1 - 1)\,k_1^{n_1}}, \tag{4}$$

where τ is the mean intensity of shear stresses of a given determination, θ_1, n_1 and k_1 are the mean temperature of the ice, the exponent, and the constant of proportionality, obtained in the determination. An averaging of the results of all available determinations yields $\bar{k} = 3.1$ kgf cm^{-2} year$^{1/3}$ degree$^{-1/3}$.

Thus, from available data, the flow law of the glacial ice is of the form :

$$\dot{\gamma}_i = -\frac{2}{\theta - 1}\left(\frac{\tau_i}{k}\right)^n \tag{5}$$

where $n = 3$, $k = 3{,}1$ kgf cm^{-2}year$^{1/3}$ degree$^{-1/3}$.

3. Stresses in a glacier

In this paper only a two-dimensional problem is discussed. An approximate solution of a three-dimensional problem by the same method does not present difficulties but it brings to much more complicated equations.

We take a rectangular system of coordinates with a horizontal axis x directed along the line of motion of the glacier from the origin or from the point on the surface in the centre of an ice cap, and an axis z directed down. We introduce the designations:

z_s and z_b — applicates of a glacier surface and its bed,

$Z = z_b - z_s$ — thickness of a glacier,

$\zeta = z - z_s$ — depth under a glacier surface,

α and β — angles of inclination of glacier surface and glacier bed to an axis x.

For determination of stresses in a glacier it is necessary to integrate differential equations of motion after the stresses on the boundaries of a glacier are determined. On the surface of a glacier the boundary conditions are:

$$\sigma_0 = -\varrho_{atm}, \; \tau_0 = 0, \tag{6}$$

since $f_0 = -\tau_0/\sigma_0 = 0$, where f—coefficient of external friction, ϱ_{atm}— atmospheric pressure, the effect of which we shall neglect. It is not difficult for the underwater surface to put in hydrostatic pressure of water, but we shall not consider that case.

On the bed of a glacier the boundary conditions will be:

$$\sigma_b = -p_b + \sigma_b', \quad \pm \tau_b = -\sigma_b f. \qquad (7)$$

Two cases are possible:

1) the bottom layer of ice is immovable, and f is a coefficient of static friction which does not characterize properties of a boundary between a glacier and its bed because shear stress acting in bottom layer is less than possible resistance;

2) the bottom layer of ice slides along the bed and f is a coefficient of dynamic friction, depending on the properties of the lower boundary of a glacier and on the velocity of sliding.

Neglecting forces of inertia, what is quite acceptable under velocities of motion which are observed in glaciers, for plane strain in the adopted coordinate system we have following equations:

$$\left.\begin{aligned} \frac{\partial \sigma_x}{\partial x} + \frac{\partial \tau_{zx}}{\partial z} &= -\varrho g \cos^2\beta \operatorname{tg}\beta \\ \frac{\partial \sigma_z}{\partial z} + \frac{\partial \tau_{zx}}{\partial x} &= -\varrho g \cos^2\beta. \end{aligned}\right\} \qquad (8)$$

Integrating in respect to z and neglecting the member $\dfrac{\partial \tau_{xz}}{\partial x}$, which is an infinitesimal quantity of higher order (*) we get:

$$\left.\begin{aligned} \frac{\partial F_x}{\partial x} &= -\bar{\varrho} g Z \cos^2\beta \operatorname{tg}\beta - \tau_{zx}(z_b) \\ \sigma_z(z) &= -\bar{\varrho} g \zeta \cos^2\beta, \end{aligned}\right\} \qquad (9)$$

where F_x indicates force acting along the axis x in the given vertical section of a glacier. According to (7) and (9)

$$\tau_{zx}(z_b) = \pm \bar{\varrho} g Z \cos^2\beta f \qquad (10)$$

and as a result of that

$$\frac{\partial F_x}{\partial x} = -\bar{\varrho} g Z \cos^2\beta (\operatorname{tg}\beta - f) \qquad (11)$$

To find F_x it is necessary to integrate (11) in respect to x. Since the beginning of the glacier is fixed as a rule (by the wall of the circus in mountain glaciers or by the mass of ice moving in the opposite direction in ice caps) the initial F_x is unknown, but at the end of a glacier ($x = l$) it is determined by boundary conditions. So

$$F_x = \int_x^l \bar{\varrho} g Z \cos^2\beta (\operatorname{tg}\beta - f)\, dx, \qquad (12)$$

and the mean longitudinal stress $\bar{\sigma}_x$ in the given vertical section of a glacier is:

$$\bar{\sigma}_x = \frac{F_x}{Z}. \qquad (13)$$

(*) According to (7) and (17) it is approximately equal to
$\varrho g \cos^2\beta\, [f(\operatorname{tg}\alpha - \operatorname{tg}\beta + 2\zeta \cos^2\beta \operatorname{tg}\beta\, \partial \operatorname{tg}\beta/\partial x) - \zeta \partial f/\partial x]$.

The mean stress in the direction of z according to (9) is:

$$\bar{\sigma}_z = \frac{1}{Z}\int_{z_s}^{z_b} \sigma_z(z)\,dz = -\frac{1}{2}\bar{\varrho} g Z \cos^2\beta\,. \tag{14}$$

From (3), (13), (12) and (14) we get mean stress deviators $\bar{\sigma}'_x = -\bar{\sigma}'_z$ in the given section of a glacier:

$$\bar{\sigma}'_x = \frac{1}{2}\left[\frac{1}{2}\bar{\varrho} g Z \cos^2\beta + \frac{1}{Z}\int_x^l \bar{\varrho} g Z \cos^2\beta(\mathrm{tg}\beta - f)\,dx\right] \tag{15}$$

In the vertical section of a glacier an absolute value of stress deviator σ'_x apparantly must decrease down under the influence of the force of friction along the bed. This alteration is interconnected with the changes of the shear stress along x and z by the condition of continuity

$$\frac{\partial^2 \dot{\varepsilon}_x}{\partial z^2} + \frac{\partial^2 \dot{\varepsilon}_z}{\partial x^2} = \frac{\partial^2 \dot{\gamma}}{\partial x \partial z} \tag{16}$$

and by equations (1), (2) and $\dot{\varepsilon}_x = -\dot{\varepsilon}_z$. In the first approximation the change of τ_{zx} along z can be considered as the linear, and proportional to the depth under the surface of a glacier, as the majority of the previous hypotheses have supposed:

$$\tau_{zx}(z) = \pm\bar{\varrho} g \zeta \cos^2\beta f \tag{17}$$

The following approximations for $\tau_{zx}(z)$ can be obtained by the solution of the systems of the equations given above.

4. The velocity of glacier motion

The horizontal components of the mean velocity of motion $\bar{u}\,(x)$ of a glacier is equal to the sum of the horizontal components of the rates of longitudinal tensions and compression $\bar{\dot{\varepsilon}}_x dx$, experienced by the ice over the entire length of the glacier from its beginning to the given point x:

$$\bar{u}(x) = \int_0^x \bar{\dot{\varepsilon}}_x(x)\,dx\,, \tag{18}$$

or taking into account (2), (5) and (15):

$$\bar{u}(x) = -\int_0^x \frac{1}{\bar{\theta}-1}\left(\frac{1}{2k}\right)^n \left[\frac{1}{2}\bar{\varrho} g Z \cos^2\beta + \int_x^l \bar{\varrho} g Z \cos^2\beta(\mathrm{tg}\beta - f)\,dx\right]^n dx, \tag{19}$$

where $\bar{\theta}$—mean temperature of ice in the vertical section of a glacier. It is possible to determine analytically the mean velocity of motion of a glacier from the equation (19) only when we know the values of variables invloved as functions of x.

The deviations of the velocities at different depths from the mean velocity are due to changes of $\dot{\varepsilon}_x$ along the vertical profile. In the first approximation alterations in the velocities of motion along the vertical line can be considered as equal to the latter of the laminar flow which is a component of complex motion of a glacier. The velocity u_{lam} of this laminar component is determined by means of integrating

the rates of shear strains along the vertical profile from the level under consideration to the bed:

$$u_{\text{lam}}(z) = \int_{z}^{z_b} \dot{\gamma}_{zx}(z)\,dz\,, \tag{20}$$

or if we neglect the influence of differences in temperature of ice at different depths:

$$u_{\text{lam}}(z) = -\frac{1}{n+1}\,\frac{2}{\theta - 1}\left(\frac{\varrho g}{k}\cos^2\beta f\right)^n (Z^{n+1} - \zeta^{n+1}). \tag{21}$$

Under the same condition the mean velocity of laminar flow is equal to:

$$\bar{u}_{\text{lam}} = \frac{1}{Z}\int_{z_s}^{z_b} u_{\text{lam}}(z)\,dz = -\frac{1}{n+2}\,\frac{2}{\bar{\theta} - 1}\left(\frac{\bar{\varrho} g}{k}\cos^2\beta f\right)^n Z^{n+1}. \tag{22}$$

Consequently, the horizontal component of the velocity of sliding of a glacier along the bed $u\,(x, Z)$ or u_{bl}, that is, the block component of motion is:

$$u_{bl}(x) = \bar{u}(x) - \bar{u}_{\text{lam}}(x) = \bar{u}(x) + \frac{1}{n+2}\,\frac{2}{\bar{\theta} - 1}\left(\frac{\bar{\varrho} g}{k}\cos^2\beta f\right)^n Z^{n+1}. \tag{23}$$

The horizontal component of the total velocity of motion of a glacier at a depth z is:

$$u(x,z) = u_{bl}(x) + u_{\text{lam}}(x,z) = \bar{u}(x)\gamma - \frac{1}{n+1}\,\frac{2}{\bar{\theta} - 1}\left(\frac{\bar{\varrho} g}{k}\cos^2\beta f\right)^n \left(\frac{1}{n+2} Z^{n+1} - \zeta^{n+1}\right) \tag{24}$$

By means of differentiation of the horizontal velocities with respect to x, we determine in the first approximation the horizontal and vertical strain rates of the ice at any depth under the surface $\dot{\varepsilon}_x(x, z) = -\dot{\varepsilon}_z(x, z)$.

The vertical component $w(x, z)$ of the velocity of motion of a glacier at a depth z is the algebraic sum of the vertical velocity of displacement in the case of motion parallel to the bed at the expense of inclination of the latter, and the integral of the vertical strain rates of the ice from the given point to the bed (*):

$$w(x,z) = u(x,z)\,\mathrm{tg}\beta - \int_{z}^{z_b} \dot{\varepsilon}_z(x,z)\,dz = u(x,z)\,\mathrm{tg}\beta + \bar{\dot{\varepsilon}}_x(Z - \zeta) -$$

$$-\frac{2}{\bar{\theta} - 1}\left(\frac{\bar{\varrho} g}{k}\cos^2\beta f\right)^n \left\{\left[\frac{1}{(n+1)(n+2)} Z^{n+1} + \zeta\left(\frac{1}{n+2} Z^n - \frac{1}{n+1}\zeta^n\right)\right]\right.$$

$$\left.(\mathrm{tg}\alpha - \mathrm{tg}\beta) - \frac{n}{(n+1)(n+2)}\zeta(Z^{n+1} - \zeta^{n+1})\left(\frac{\partial f/\partial x}{f} - 2\cos^2\beta\,\mathrm{tg}\beta\,\frac{\partial\,\mathrm{tg}\beta}{\partial x}\right)\right\}. \tag{25}$$

(*) The actual velocity of vertical motion differs from that described by equation (25) also due to compaction of the ice with depth. The method for accounting for this factor is given in a paper by the author (1959), English translation (1960).

From (25) the vertical velocity component at the surface of the glacier $w(x.z_s)$ is equal to:

$$W(x) = \bar{u}(x)\mathrm{tg}\beta + \bar{\dot{\varepsilon}}_x Z - \frac{1}{(n+1)\,(n+2)}\,\frac{2}{\theta - 1}\left(\frac{\bar{\varrho} g}{k}\cos^2\beta f\right)^n Z^{n+1}\mathrm{tg}\alpha, \tag{26}$$

and for velocity at the bed $w(x, z_b)$ the second and the third terms converts to zero. The mean vertical velocity of the vertical component of motion is:

$$\bar{w}(x) = \frac{1}{Z}\int_{z_s}^{z_b} w(x,z)\,dz = \bar{u}(x)\mathrm{tg}\beta + \frac{1}{2}\bar{\dot{\varepsilon}}_x Z -$$

$$- \frac{1}{(n+2)}\,\frac{1}{\bar{\theta} - 1}\left(\frac{\varrho g}{k}\cos^2\beta f\right)^n Z^{n+1}$$

$$\left[(\mathrm{tg}\alpha - \mathrm{tg}\beta) + \frac{n}{n+3} Z\left(\frac{\partial f/\partial x}{f} - 2\cos^2\beta\,\mathrm{tg}\beta\,\frac{\partial\,\mathrm{tg}\beta}{\partial x}\right)\right]. \tag{27}$$

More precise solutions can be obtained by the method of successive approximation from the system of equations given above which includes the condition of continuity (16).

From equations (18) to (27) it follows that the velocity of motion at any point in a glacier depends not only upon the conditions in the given vertical section of the glacier, as was supposed by the theory of laminar flow, but also on the shape of the surface, the bed, the thickness and changes in the coefficient of friction against the bed throughout the length of the glacier. However, the integral equations obtained may be transformed into differential equations expressing the relationship between all the characteristics of the given cross-section of the glacier and providing extensive opportunities for analysis of the relationship of dynamics, morphology, climatic conditions and geological activities of glaciers. This has been done, in part, in the author's paper «The Dynamics and Morthology of Glaciers».

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GLACIOLOGICAL STUDIES ON THE ROSS ICE SHELF

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SUMMARY

Parallel crevasses oriented normal to the axes of firn folds in the Ross Ice Shelf between Roosevelt Island and the Bay of Whales, Antarctica, were studied during the IGY 1957-58 and 1958-59 seasons. Seismic soundings reveal that the ice shelf ranges from 70 m to 133 m in thickness in the area of intense folding and heavy crevassing, and is floating in sea water mores than 500 m deep. The folds and associated crevasses are produced by horizontal compression stresses whidh are induced by merging of two ice streams after they flow around the east and west sides of Roosevelt Island. The density-depth profile of the upper 15 m of firn in the folded area reveals higher densities than those found at the same depths in other cold glaciers of the world. The increased densification in the folded areas of the shelf ice is attributed to plastic deformation resulting from lateral compressive stresses.

The difference in accumulation rate between the crest and flank of one fold provided the key for determining the age of that fold through firn stratigraphy. By this method, a fold with a wave length of 140 m and amplitude of 10 m was sufficiently well developed in 1926 (33 years before this study) so that the crest received less accumulation because of its increased height above the surrounding area.

Strain rates show that the axes of the firn folds are almost always perpendicular to the principal compressive axis, and that the crevasses are essentially perpendicular to the principal tension axis. An exception exists in the vicinity of the anticlinal crests where the crevasses make a considerable angle with the principal stress axes. A comparison of strain rates in crevassed and non-crevassed areas defines a rupture criteria which takes the mathematical form of a surface of revolution around the line $s_1 = s_2$. An estimate of the anticline's age based on srtain data gives a value of 20-30 years.

A thin ice layer in the study anticline was also deformed and displayed a moderately strong preferred crystal orientation. The maximum concentrations of the c-axes is 8 % to 9 % per one per cent of the area. Fabric diagrams plotted on a Schmidt equal-area net generally show four maxima 12° to 17° from the equator, and inclined 20° from one of the planes of principal shear but essentially normal to the other shear plane. The crystal fabric of a ptygmatic fold on one of the anticlinal limbs suggests a mechanism of deformation involving differential slip along the basal plane which is nearly parallel to the axial plane of the anticline.

THERMAL AND DYNAMIC GLACIAL REGIMES

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SUMMARY

The peculiarities of glacial development are in large measure determined by the thermal and dynamic regimes of the glacier. The physical processes underlying these regimes are interrelated, and their study requires a joint consideration of phenomena with account taken of their mutual influence.

The temperature field of a glacier is the result of the effects of many factors. It depends on the thermo-physical properties of the snow and ice and upon the conditions on the surface of the glacier and at the point of contact with the underlying. bedrock. To a large extent, the thermal regimen of a glacier is affected by ice movement and the accumulation of snow on its surface. Ice movement is associated with the transport of heat and with heat generation in the interior due to the transformation of mechanical energy into thermal energy. Accumulation of mass can alter the position of the surface; gradual settling of the snow and its compaction are likewise associated with the transfer of heat. Also possible are the effects of phase transformations and the the movement of meltwater.

The temperature field of a glacier is described by an intricate differential equation of the Fourier-Kirkhoff type with additional terms that account for the transport of heat and for supplementary heat generation. The boundary conditions are specified on the surface, the position of which (in the general case) can vary in time, and at the base.

The temperature fields are obtained with the aid of a hydraulic integrator and the analytical method. A solution of the problem has been obtained as applicable to the Antarctic Ice Sheet. A comparison of the solutions with the data of field observations has shown satisfactory coincidence.

The dynamic regimen of a glacier is determined, on the one hand, by that fact that, due to its viscous-plastic properties, the ice flows under the action of gravitational forces, and, on the other hand, by the accumulation of mass on its surface.

As a result of subsidence and compaction, the snow accumulated on the surface, is transformed into ice, thus forming the body of the glacier. Variations in the rate of accumulation can cause the formation of a glacier, lead to more intense growth, or, on the contrary, it can make the glacier retreat, and, gradually, even lead to its total disappearance. Indirectly, these processes are affected by the temperature conditions, and, consequently, by variations in climate that are connected with the life of the glacier itself.

The formative processes of a glacier and its flow are, to some extent, similar to the processes of formation of filtering areas on the ground and to the filtration of ground water deep underground. This analogy has made it possible to take advantage of the well-known mathematical apparatus of the theory of nonctationary filtration for describing the dynamic regimen of glacier.

In view of the absence of a complete analogy, the peculiarities of the phenomenon are taken into consideration. The principal peculiarity is that the spreading of the ice follows a slightly more complicated law than that of filtration. An exceedingly convenient feature is that in the proposed formulation, one can also solve the dynamic problem by means of the hydraulic integrator, wich opens up broad possibilities and enables one to bring the calculation scheme as close as possible to real conditions. Several calculations have been carried out dealing with the formation and retreating of a glacier as exemplified by the Antarctic Ice Sheet.

The problem of subsequent calculations consists in determining to what extent the existing configuration of the glacier differs from an earlier one, and in expressing suppositions concerning the period of formation of the glacier and the possible time of its retreat.

A certain complication in the calculation scheme will allow for a combined consideration of the thermal and dynamic regimes under conditions of complete mutual influence, with allowance made for these processes being non-steady.

THE DYNAMICS AND MORPHOLOGY OF GLACIERS

SHUMSKIY P. A. (U.S.S.R.)

1. Introduction

The existing quantitative theories of glacier movement are divided into two groups: *a)* kinematic and *b)* dynamic.

The kinematic theories, which are based on balance methods and the condition of continuity, establish a relationship between motion, intensity of accumulation, ablation, dimensions and shape of the glacier. To this category belong methods of calculating the movement from the mass balance, that have been known since last century, including the most refined theory of S. Finsterwalder (1897).

A particular place is occupied by a method—proposed in 1958 by V.N. Bogoslovskiy— of calculating the motion from the heat balance of the glacier; this method combines the kinematic approach, when considering processes of the advective transfer of heat, with the dynamic approach, when accounting for the internal sources of heat generation.

The dynamic theories regard motion as a function of the physico-mechanical properties of the ice, the stresses and strains arising in it, which are likewise conditioned by the shape and dimensions of the glacier.

The joint application of the balance kinematic and dynamic methods makes it possible to analyze the dependence of the state and evolution of glaciers upon the terrain and heat-and mass-exchange with the surrounding medium.

Up till now, the dynamic theories have not been able to account for the chief element in glacier motion -block sliding along the bed, with which the geological action of the glacier is associated. The author has elaborated a new dynamic theory, which to a certain extent eliminates this deficiency (*). The present paper is devoted to an analysis of certain relationships between the conditions of existence of glaciers, their dynamics and morphology on the basis of the proposed theory.

A differential equation of the state of a glacier

According to ([1]), the horizontal component u (x) of the average velocity of motion in a given vertical profile of a glacier at a distance x from its beginning (or the centre of the ice cap) along the line of motion under conditions of deformation only in the plane xz $(\sigma_y' = 0, \dot{\varepsilon}_y = 0)$, is equal to

$$\bar{u}(x) = \int_0^x \overline{\dot{\varepsilon}_x}(x)\,dx, \tag{1}$$

where the horizontal component $\overline{\dot{\varepsilon}_x}$ of the mean relative rate of longitudinal strain is:

$$\overline{\dot{\varepsilon}_x} = -\frac{1}{\bar{\theta} - 1}\left(\frac{\bar{\sigma}_x'}{k}\right)^n. \tag{2}$$

(*) See, in this book, the paper «On the Theory of Glacier Motion». Hereafter, references to this paper will be designated by ([1]).

In its turn, horizontal component of the mean deviator of the stress σ_x' is defined as:

$$\sigma_x' = \frac{\mathscr{F}_x'}{\mathscr{L}}, \tag{3}$$

where the horizontal component F_x' of excess longitudinal force, acting in the given vertical cross section of the glacier, over the force of friction along the bed is equal to:

$$\mathscr{F}_x' = \frac{1}{2}\left[\frac{1}{2}\bar{\varrho} g \mathscr{L}^2 \cos^2\beta + \int_x^l \bar{\varrho} g \mathscr{L} \cos^2\beta (\operatorname{tg}\beta - f)\, dx\right] \tag{4}$$

Equating the values of the gradient $\frac{\partial \mathscr{F}_x'}{\partial x}$ of change of the longitudinal force acting in the glacier, determined on the basis of glacier characteristics from (4) and on the basis of the rate of strain from (2) and (3), we obtain the differential equation:

$$\bar{\varrho} g \cos^2\beta \left(\operatorname{tg}\alpha + \mathscr{L} \cos^2\beta \operatorname{tg}\beta \frac{\partial \operatorname{tg}\beta}{\partial x} - f\right) =$$

$$= -2(\bar{\theta} - 1)^{1/n} k \bar{\dot{\varepsilon}}_x^{1/n} \left(\frac{\operatorname{tg}\alpha - \operatorname{tg}\beta}{\mathscr{L}} - \frac{1}{n} \frac{\partial \bar{\dot{\varepsilon}}_x / \partial x}{\bar{\dot{\varepsilon}}_x}\right), \tag{5}$$

that relates the physico-mechanical properties and the temperature of ice, the gravity acceleration, the thickness, shape, rate of strain, and the coefficient of friction of a glacier against its bed in a given vertical cross section.

Equations (1) to (4) and (5) permit examining the relationship between the principal morphological and dynamic characteristics of the glacier.

Let us first consider the friction of the glacier against its bed, the role of which was not taken into account by the majority of earlier theories.

3. Friction of a glacier against its bed

The force of friction of the glacier against its bed produces a considerable effect on the motion and shape of the glacier. The greater is the force of friction in the lower-lying part of the glacier as compared with the moving force, or in other words, the greater is the coefficient of friction f as compared with the angle of inclination of the surface α in the direction of motion of the glacier, the lower is the velocity of motion at the given point of the glacier, and, according to equations (22) and (23) ([1]), the greater the role played by laminar flow and the smaller the part played by block sliding in the motion of the glacier. Inasmuch as the product of the mean velocity of motion in the given vertical cross section of the glacier by its thickness is limited by the mass balance of the upperlying part of the glacier (equation (7)), the force of friction against the bed affects the shape and thickness of the glacier: the higher is the coefficient of friction at the given point of the glacier, the steeper is the inclination of its surface in the direction of motion, and, hence, the greater is the coefficient of friction in the lower-lying part of the glacier, the greater is the thickness of the glacier at the given point.

Up till now, the magnitude of the force of friction could not be determined, and the regularities of its variation remained unknow. The only obvious thing is that it should depend on: 1) the physical properties and roughness of the glacier bed; 2) the temperature of the bottom layer (which follows from (5)), 3) the velocity of

sliding, and 4) the quantity and properties of the morainic material in the bottom layer of ice. The influence of the first two factors is very clearly manifested in the regions of sheet glaciation.

In floating ice shelfs the bottom friction is practically zero, and so, with a comparatively small thickness and negligible surface inclination they possess a high velocity of motion that is exclusively due to the longitudinal tension without any shearing of the upper layers over the lower ones. But when the glacier begins to rest on a solid bed and the hydrostatic pressure of the water is not able to relieve considerably the sharply increasing force of the bottom friction, we encounter a steep slope of the ice cap and a sharp increase in the thickness of the glacier.

The gradual diminishing surface inclination from edge to centre in the caps themselves is, apparently, due partly to the reduction (in this direction) of the coefficient of friction due to increasing temperature of the bottom layer and to the appearance, on the bottom, of increasing quantities of melt-water that plays the role of a lubricant.

If we know the properties and temperature of the ice, the shape of the glacier, and the distribution of velocity of motion, it is possible to calculate the value of the force of friction from (5), or from (21) or (24) and (19) (:) (by means of successive integration upwards along the glacier by small sections, within the limits of which the variation of f with respect to x is assumed linear).

The dependence of the shape, thickness and velocity of motion of the ice cap overlying a horizontal basis, upon the value of the coefficient of friction against the bed (which dependence is given by equation (5)) is shown schematically in Fig. 1.

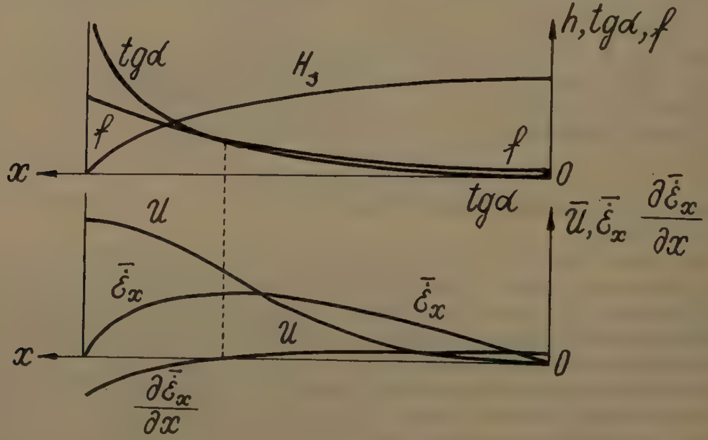

Fig. 1 — Shape (tg α), thickness ($Z = H_s$), rate of longitudinal strain ($\bar{\dot{\varepsilon}}_x$), velocity of motion ($u$), and coefficient of friction (f) of an ice cap against the bed.

Measurements of glacier motion show that it is frequently of a nonuniform, pulsating nature with considerable fluctuations in magnitude and even in direction of motion over short intervals of time. These fluctuations cannot be explained either by changes in the mass balance and shape of the glacier or by temperature fluctuations of the ice. An account not only of local tangential stresses but also of normal stresses, as carried out in ([1]), leads to the conclusion that, due to transmission of pressure in the glacier, on sections with a high coefficient of friction (that stems from the excessive roughness of the bed or the abundance of morainic material in the bottom layer),

tangential stresses at the bed can attain a very great magnitude. In certain conditions, this, apparently, leads to the melting of the bottom layer of ice and the formation of socalled «blue bands» (Shumskiy, P.A., 1955, 1958). This should be accompanied by a sharp reduction in the force of friction in the bottom layer. In all probability, periodic decreases and increases in the coefficient of friction in the case of ice melting along the planes of scaly thrust-faults in the bottom layer of the glacier and subsequent crystallization of the melt-water are the cause of pulsations in glacier motion.

One of the most important tasks of glaciology is to make a study of the laws of variation of the force of friction of glaciers against the bed under different conditions.

4. Inclination of the surface of a glacier

From (5) it follows that inclination of the surface of a glacier depends on a whole series of factors.

Firstly, the surface inclination of a glacier depends on the inclination of its bed-not directly, but as a function of the product of the angle of inclination of the bed by the rate of longitudinal strain of the glacier. The sign and manigtude of the latter by themselves condition, to a certain extent (inasmuch as this is not hindered by the processes of accumulation or ablation on the surface), a definite relation between the angles of inclination of the bed and the surface, so that not every combination of variations of these quantities is possible. For this reason, in different part of the glacier the variations in surface inclination repeat, in general, the variations of the inclination of its bed, but with certain deviations. The surface inclination in the direction of motion becomes greater than the bed inclination in the case of longitudinal tension, and less than it in the case of compression, and the magnitude of this deviation is directly proportional to the roots of the nth power of the mean rate of longitudinal strain of the glacier and its negative temperature, and inversely proportional to the thickness, density, fluidity of ice and the gravity acceleration.

Secondly, the surface inclination of a glacier depends upon the curvature of the bed: it is the greater, the more is the convexity of the bed when the latter is inclined in the direction of motion, or the concavity of the bed in the case of inverse inclination.

Thirdly, the surface inclination is a function of the gradient of changes of longitudinal strain rates of a glacier in the direction of motion: it is the greater, the smaller is the gradient of growth of the rate of longitudinal tension, or the greater is the gradient of growth of the rate of longitudinal compression in the direction of motion of the glacier. In this relationship, the factor of proportionality is the longitudinal strain rate of ice, its temperature, fluidity, density, and the gravity acceleration.

Finally, as has been pointed out above, the surface inclination of a glacier in the direction of motion is the greater, the higher is the coefficient of friction of a glacier against its bed.

Under certain combinations of the above-enumerated conditions, movement of the entire mass of ice along an uneven bed becomes impossible. Still, if the glacier has the surface shape and distribution of strain rates which do not correspond to a given bed profile and coefficient of bottom friction, this is a result of the fact that the sliding is not along the bed, but along the planes of the thrust faults that cut off fixed masses of ice lying in depressions, in front of the projections of the bed, and/or behind them.

5. Steady state of a glacier

A glacier is in a state of dynamic equilibrium on the condition that the mass

balance of each of its parts is maintained. The equation of the mass balance of a cross-section of a glacier may be written in general form as follows.

$$\bar{\varrho} S \bar{u} = \mathscr{A}, \tag{6}$$

where S is the area of the cross-section, A is the mass balance on the upper and lower boundaries of the glacier above the given cross section. For conditions of a plane problem, (6) assumes the form:

$$\bar{\varrho} \mathscr{Z} \bar{u} = \mathscr{A}, \tag{7}$$

$$\mathscr{A} = \int_0^x a(x)\,dx = \int_l^x a(x)\,dx, \tag{8}$$

where a is the mass balance on the upper and lower boundaries of the given elementary column of ice of dimensions $Z\,dx\,dy$.

Differentiating (7) with respect to x, we obtain:

$$\frac{a}{\bar{\varrho}} + \bar{u}(\mathrm{tg}\alpha - \mathrm{tg}\beta) - \bar{\dot{\varepsilon}}_x Z = 0 \tag{9}$$

From (9) we find the value of the mean rate $\bar{\dot{\varepsilon}}_x$ of strain of the glacier in the direction x, and then its derivative $\frac{\partial \bar{\dot{\varepsilon}}_x}{\partial x}$.

Substituting the values found into (5), we get :

$$\bar{\varrho} g \cos^2\beta\left(\mathrm{tg}\alpha + Z\cos^2\beta\,\mathrm{tg}\beta\,\frac{\partial \mathrm{tg}\beta}{\partial x} - f\right) = -$$

$$(\theta - 1)^{1/n}\,\frac{k}{n}\,\frac{2}{Z^{1/n}}\left[\frac{a}{\bar{\varrho}} + \bar{u}(\mathrm{tg}\alpha - \mathrm{tg}\beta)\right]^{1/n}\left[\frac{\frac{e}{\bar{\varrho}}\,\mathrm{tg}\alpha - \bar{u}\left(\frac{\partial \mathrm{tg}\alpha}{\partial x} - \frac{\partial \mathrm{tg}\beta}{\partial x}\right)}{\frac{a}{\bar{\varrho}} + \bar{u}(\mathrm{tg}\alpha - \mathrm{tg}\beta)} + (n-2)\,\frac{\mathrm{tg}\alpha - \mathrm{tg}\beta}{Z}\right] \tag{10}$$

where $e = -\frac{\partial a}{\partial z}$ is the gradient of increase of the rate of net accumulation with altitude («glacierization energy», P.A. Shumskiy, 1947).

Equations (7) to (10) relate the physico-mechanical properties and temperature of the ice, the gravity acceleration, the coefficient of friction against the bed, the thickness, shape, velocity of motion, and of accumulation—ablation processes of a steady glacier, and make it possible to consider the dependence of its principal morphological and dynamic characteristics upon the mass-exchange on its upper and lower boundaries. We shall dwell only on the most important characteristics.

In accordance with (7), the thickness and velocity of motion of a steady glacier at a given point are proportional to the intensity of feeding of its upper-lying part. At the same time, from (9) it follows that the thickness of a glacier is directly proportional to the intensity at the given point and to the product of the velocity of motion by the magnitude of the surface inclination with respect to the bed in the direction of motion, and inversely proportional to the rate of longitudinal tension of a glacier. In the region of ablation, in the case of longitudinal compression of a glacier, the

numerator and denominator of the fraction characterizing its thickness convert into negative values, and in the case of continuing longitudinal tension the surface inclination with respect to the bed increases to such an extent that the numerator remains positive, despite the ablation of the surface.

If we solve (10) in respect to $\bar{u}$ we can see, that the rate of motion of a steady glacier is directly proportional to «glacierization energy», and to the intensity of feeding at the given point.

6. The role of the relief of the underlying surface and of climate conditions in the dynamics of glaciers

By means of differentiating the equation (4) in respect to x it is possible to show that the force which brings a glacier into motion consists of two parts:

1) the run-off force, which is due to inclination β of the glacier bed ($\bar{\varrho} g Z \cos^2 \beta \operatorname{tg}\beta$), and

2) the out-flow force, which is due mainly to inclination $\alpha - \beta$ of the surface of the glacier with respect to the bed

$$(\bar{\varrho} g Z \cos^2\beta\, [\operatorname{tg}\alpha - \operatorname{tg}\beta + Z \cos^2\beta \operatorname{tg}\beta\, \partial \operatorname{tg}\beta / \partial x]).$$

Substituting, in equation (10), $(\operatorname{tg}\alpha - \operatorname{tg}\beta) + \operatorname{tg}\beta$ for $\operatorname{tg}\alpha$ and solving it for $\operatorname{tg}\alpha - \operatorname{tg}\beta$, it may be shown that the magnitude of the surface inclination relative to the bed in the direction of motion in a steady glacier is directly proportional to the «glacierization energy» and intensity of feeding at the given point. Such an inclination arises in places where equilibrium between accumulation and outflow of ice (or ablation and its inflow) cannot be achieved by the run-off force alone (due to inclination of the glacier bed). Consequently, the ratio of the out-flow force to the entire moving force is a quantitative expression of the role of the climatic factor in glacier motion, whereas the ratio of the run-off force to the entire moving force is a quantitative expression of the role of the relief of the underlying surface. These characteristics may be obtained both for a single point or part of the glacier and for the glacier as a whole.

Differentiation with respect to x of the equation of the horizontal velocity of block sliding (equation (23) in ([1])) yields:

$$\dot{\varepsilon}_{xbl} = \bar{\dot{\varepsilon}}_x - \frac{2}{\bar{\theta} - 1}\left(\frac{\varrho g}{k}\, \mathscr{Z} \cos^2\beta f\right)^n \left[\frac{n+1}{n+2}(\operatorname{tg}\alpha - \operatorname{tg}\beta) - \right.$$

$$\left. - \frac{n}{n+2}\, Z \left(\frac{\partial f/\partial x}{f} - 2\cos^2\beta \operatorname{tg}\beta\, \frac{\partial \operatorname{tg}\beta}{\partial x}\right)\right]. \qquad (11)$$

From (11) it is seen that the greater is the surface inclination of a glacier with respect to the bed in the direction of motion and the convexity of the bed inclined in the direction of motion, the greater is the acceleration, and, hence, the velocity of block sliding of the glacier for one and the same coefficient of friction against the bed. Together with increasing rate of block sliding, there is a fall in the portion of the laminar component of motion.

Thus, block sliding of a glacier along the bed is caused mainly by the out-flow force and is proportional to the «glacierization energy» and the intensity of feeding of the glacier, that is, by its very nature it reflects the role of the climatic factor in the motion of a glacier. On the contrary, laminar flow is caused mainly by the run-off force, and quantitatively reflects the role of the bed relief in the motion of a glacier.

7. A MORPHOLOGICAL-DYNAMIC CLASSIFICATION OF GLACIERS

The above-characterized mutual relationships between the conditions of existence of glaciers, their morphology and dynamics permit us to regard the relationship of run-off and out-flow forces as an objective and precise basis for the classification of glacial formations.

Since this basis is quantitative, it admits any degree of subdivision. We shall confine ourselves to an enumeration of only the chief types of glaciers.

1. *Run-off glaciers* or mountain glaciers. The total run-off force of the entire glacier comprises 100% of the moving force (the total out-flow force is equal to zero or is negative). Predominant or important is the role of laminar flow in the motion. Morphologically and dynamically, the run-off glaciers are totally governed by the relief of the terrain, they occupy depressions on the slopes of mountainous elevations and, in plan, are extremely cut up; the direction of their motion coincides, in general, with the direction of inclination of the bed; they have irregular surface shapes that duplicate the irregularities of the bed. In the accumulation region, the surface is concave, the thickness increases downwards; a regular convex form is encountered only in the region of ablation. In the case of all-round development, in plan, run-off glaciers are represented by thin covers (on regular volcanic cones).

2. *Montain-sheet glaciers* or glaciers with predominant run-off. The total run-off force is from 50 to 100% of the moving force. The motion is laminar-block. Morphologically, these are saddle-type glaciers moving in opposite directions away from the watershed of the mountain; and in the cases of all-round development in plan, they are irregular sheets, relatively thin on elevations and thick in valleys and depressions.

3. *Sheet glaciers* or glaciers with predominant out-flow. The total out-flow force amounts to 50-100% of the moving force. The motion is block-laminar. In mountainous countries, this category is represented by glaciation of the «Spitzbergen type», when in less divided countries by the ice sheets with bulges—out-flow centres on elevations of the bed.

4. *Out-flow glaciers*. The total out-flow force is 100% of the moving force (the total force of run-off is equal to zero or is negative). The motion is almost completely or completely represented by block sliding; this is aided by the distribution of temperature in the ice mass (low temperature of the main mass with a relatively warm bottom layer in the ice caps). To this category belong floating ice shelfs and ice caps, with centres or ice-divides above bed depressions, the shape and direction of motion of which are completely independent of the terrain. The transformation of an ice sheet into an ice cap is associated with migration of the centre from the elevation of the bed to its depression or with isostatic subsidence of the earth's crust in the central part of the ice cap.

Despite the common features in the internal laws of development of glacial phenomena, the morphology and dynamics of run-off glaciers is determined solely by the relief of a terrain and is independent of the climatic conditions (latitudinal climatic zones), whereas out-flow glaciers are characterized by total independence from the terrain and are governed only by the climatic conditions, or more precisely, by the conditions of energy-mass-exchange with the surrounding medium.

8. Intensity of geological activity of glaciers

Action of glaciers on the earth's crust embodies three mutually related aspects: erosional, transport, and accumulative activity. The intensity of the accumulative activity is determined by the intensity of transportation of morainic material and by the rate of ablation of ice. The rate of ablation, which depends on the heat balance on the glacier boundaries, determines only the distribution of deposited morainic material; and, on the whole, the intensity of geological work of a glacier is dependent on the intensity of the processes of erosion of the bed and transportation of the morainic material.

It we put aside the well-known processes of transportation of material brought onto a glacier from above, the intensity of action of a glacier on the bed (the quantity of rock removed from unit area in unit time) will be caused by the intensity of erosional activity (the rate of the destruction of rocks), and by the possible intensity of denudational activity (the possible rate of removal of the products of erosion).

The intensity of erosional activity depends, on the one hand, on the physico-mechanical properties of the ice and the bed rocks, which determine the degree of their stability, and on the temperature regime of the bottom layer, which determines the possibility of participation of the processes of weathering, and, on the other hand, upon the forces of friction of the glacier against its bed, that is, on the magnitudes of ice pressure and of the coefficient of bottom friction.

The possible intensity of denudational activity of a glacier is directly proportional to the velocity of motion of the bottom layer of ice, that is, on the velocity of block sliding of a glacier along the bed. This latter, as follows from equations (23) and (19) in ([1]), is a function of the difference between the moving force and the force of friction of the lower part of a glacier.

Consequently, the intensity of geological activity of a glacier is in two ways dependent upon its force of friction against the bed: the growth in force of friction intensifies the erosional activity, but simultaneously weakens the denudational activity of a glacier. The summary result reaches a maximum for a definite optimum magnitude of the force of bottom friction. Thus, a glacier with only laminar flow would not at all act mechanically on the bed due to the excessively high force of bottom friction, whereas a glacier that only slides along its bed does not produce any mechanical action on it due to the absence of bottom friction (the floating ice shelfs are an example).

For a given coefficient of friction against the bed, the magnitude of which is conditioned by the properties and temperature of the bottom layer, a glacier will act on the bed the more strongly, the higher is its thickness and the velocity of sliding along the bed. As has been shown above, these two values, other conditions being equal, are greater in the case of out-flow glaciers than in the case of run-off glaciers. Thus, the intensity of the geological action of glaciers is directly proportional to the magnitude of the relative role of the climatic factor in the motion of glaciers. Other conditions being equal, subjected to the terrain run-off glaciers act on the bed less actively than the out-flow glaciers, the dynamics and morphology of which are governed by climatic conditions.

9. The evolution of glaciers

Alterations in the energy-mass-exchange with the surrounding medium, alterations in relief of the terrain, sea level, etc. (which are partially caused by the glaciers themselves) and also certain internal changes, for example, in the coefficient of bottom friction, are constantly bringing glaciers out of the steady state, while the most part of internal processes in glaciers are directed towards restoration of the deranged

equilibrium. The equations of the mass balance for unsteady glacier can be written in the form:

$$\int_0^x \left(\frac{a}{\varrho} - \frac{\partial Z}{\partial t}\right) dx = \bar{u}\ Z = \int_l^x \left(\frac{a}{\varrho} - \frac{\partial Z}{\partial t}\right) dx\,, \tag{12}$$

where t—time, and differentiating in respect to x gives for the velocity $\frac{\partial Z}{\partial t}$ of change of glacier thickness value:

$$\frac{\partial Z}{\partial t} = \frac{a}{\varrho} + \bar{u}(\mathrm{tg}\alpha - \mathrm{tg}\beta) - \bar{\dot{\varepsilon}}_x Z. \tag{13}$$

Differentiating (13) in respect to x and supposing that $\frac{\partial \mathrm{tg}\beta}{\partial t}$ 0, we find the rate of change of angle of surface inclination of a glacier:

$$\frac{\partial \mathrm{tg}\alpha}{\partial t} = \frac{e}{\varrho}\mathrm{tg}\alpha - 2\bar{\dot{\varepsilon}}_x(\mathrm{tg}\alpha - \mathrm{tg}\beta) - \bar{u}\left(\frac{\partial \mathrm{tg}\alpha}{\partial x} - \frac{\partial \mathrm{tg}\beta}{\partial x}\right) + Z\frac{\partial \bar{\dot{\varepsilon}}_x}{\partial x} \tag{14}$$

Next differentiating (12) and (13) in respect to time we have:

$$\int_0^x \left(\frac{1}{\varrho}\frac{\partial a}{\partial t} - \frac{\partial^2 Z}{\partial t^2}\right) dx = \bar{u}\frac{\partial Z}{\partial t} + Z\frac{\partial \bar{u}}{\partial t} = \int_0^x \left(\frac{1}{\varrho}\frac{\partial a}{\partial t} - \frac{\partial^2 Z}{\partial t^2}\right) dx - \left(\frac{a}{\varrho} - \frac{\partial Z}{\partial t}\right)_{l,t}\frac{dl}{dt}, \tag{15}$$

$$\frac{\partial^2 Z}{\partial t^2} = \frac{1}{\varrho}\frac{\partial a}{\partial t} + \bar{u}\frac{\partial \mathrm{tg}\alpha}{\partial t} + \frac{\partial \bar{u}}{\partial t}(\mathrm{tg}\alpha - \mathrm{tg}\beta) - Z\frac{\partial \bar{\dot{\varepsilon}}_x}{\partial t} - \bar{\dot{\varepsilon}}_x\frac{\partial Z}{\partial t}. \tag{16}$$

The equation (15) allows to determine the velocity of advance and retreat of a glacier end. At the snout of a glacier ending by a slope of final steepness the simple geometric relationship takes place:

$$\frac{dl}{dt} = \frac{\partial Z/\partial t}{\mathrm{tg}\alpha - \mathrm{tg}\beta}. \tag{17}$$

Considering that $Z(l) = 0$, from (15) and (17) we have:

$$\frac{dl}{dt} = \frac{\int_0^l \left(\frac{1}{\varrho}\frac{\partial a}{\partial t} - \frac{\partial^2 Z}{\partial t^2}\right) dx\,,}{\frac{\partial Z}{\partial t} + \bar{u}(\mathrm{tg}\alpha - \mathrm{tg}\beta) - \frac{a}{\varrho}} \tag{18}$$

where all the values out of the integral refer to the end of a glacier. Since a at the end of a glacier is a negative value, in the denominator of the right part of the equation (18) we have the sum of absolute values of rates of surface rising of a glacier end at the expense of growth of its thickness $\left(\frac{\partial Z}{\partial t}\right)$ and horizontal motion ($\bar{u}[\mathrm{tg}\alpha - \mathrm{tg}\beta]$) and a rate of surface lowering at the expense of ablation. Thus, accordingly to the equation (18) the velocity of displacement of a glacier end is directly proportional to surplus of acceleration of summary mass-exchange between a glacier and external medium

above acceleration of growing of glacier thickness and inversely proportional to approximately doubled rate of external mass-exchange at the end of a glacier. The more the velocities of motion and ablation at the end of a glacier are the less is the velocity of its displacement under the influence of one and the same change of rate of external mass-exchange.

If a glacier, coming down to water, ends by a cliff at the line of hydrostatic balance where masses $\bar{\varrho}\,dx\,dy\,Z$ of elementary column of ice and water are equal to each other, then instead of (17) the following relationship takes place:

$$\frac{dl}{dt} = \frac{\bar{\varrho}}{\bar{\varrho}\,\mathrm{tg}\alpha + (\bar{\varrho}_w - \bar{\varrho})\mathrm{tg}\beta}\,\frac{\partial Z}{\partial t} = \left(\frac{\partial \bar{u}}{\partial t} + \frac{1}{\bar{\varrho}}\frac{\partial \bar{a}_x}{\partial t}\right)dt, \tag{19}$$

where $\bar{\varrho}_w$—the mean density of water at the front of a glacier and a_x—the mean velocity of ice discharge from the frontal cliff (in horizontal direction). From (15) and (19) we obtain:

$$\frac{dl}{dt} = \frac{\displaystyle\int_0^l \left(\frac{1}{\bar{\varrho}}\frac{\partial a}{\partial t} - \frac{\partial^2 Z}{\partial t^2}\right)dx + \int_{z_s}^{z_b} \frac{1}{\bar{\varrho}}\frac{\partial a_x}{\partial t}\,dz}{\displaystyle Z\frac{1}{dt} + \frac{dZ}{dt} + \bar{u}\left(\mathrm{tg}\alpha + \frac{\bar{\varrho}_w - \bar{\varrho}}{\bar{\varrho}}\mathrm{tg}\beta\right) - \frac{a}{\bar{\varrho}}}. \tag{20}$$

Consequently under these conditions the velocity of displacement of the glacier end is proportional to surplus of acceleration of summary external mass-exchange, including now also the mass-exchange through frontal cliff, and inversely proportional to the thickness of a glacier plus approximately the doubled rate of external mass-exchange at the end of a glacier through its upper surface.

In case of a sudden change of the altitude H of the snow line on $\Delta H = H_1 - H_0$, of «energy of glacierization» on $\Delta e = e_1 - e_0$ and the resultative change of rate of accumulation on $\Delta a(x) = a_1(x) - a_0(x)$, equations (13) and (14), equations (15) and (19) from [1] differentiated in respect to time, and, at last, (16) make it possible to determine the velocity and duration of change of glacier size and regime of glacier motion up to reaching a new steady state.

On the basis of the above theory it is possible to solve different problems connected with variations of glaciers though it demands complicated and bulky calculations.

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